35. GLOBAL SEDIMENTARY MASS BALANCE AND SEA LEVEL CHANGES

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1. Introduction

It has long been thought that the sedimentary rocks of the earth can be considered the ultimate product of weathering of igneous materials which originally formed the primordial crust and mantle. Mead (1907) postulated that the composition of average sedimentary rock plus seawater must equal the composition of the igneous rock from which they were derived. Plate tectonics and the possibility of loss through subduction presents the modern geologist with a more complex system. Weathering leaches out certain elements that are easily removed through aqueous solution and leaves a residue of materials that are solution resistant and remain behind as soil. Rivers are the major agent of transport of the dissolved and particulate materials bringing them to the sea. At the river mouths, the materials may be deposited on the continental shelves and thus remain part of the continent proper, or may bypass the shelf and enter the ocean basin. Much particulate sediment is deposited on the continental slopes and rises, but substantial quantities of solutes are fixed by organisms in the mixed surface layer of the ocean to later rain down on the ocean floor and form pelagic sediments. Sediments and ocean crust are subducted beneath oceanic trenches in the Benioff zones, which carry these differentiates to depths where they melt and may be returned to the continents as volcanics. At Benioff zones, it is also possible that minor amounts of oceanic sediment may be scraped off and incorporated directly into active continental margins. At certain times during the history of the Earth, continents have collided, with the result that continental shelf, slope, rise, and abyssal plain deposits and even fragments of the ocean crust have become incorporated into mountain ranges.

These processes, physical, chemical, and biologic in nature, operate in a complex way and on very different time scales. The ultimate control on sedimentation, the rates at which igneous and sedimentary materials can be weathered to form solutes and particles that can be transported, may be a function of the composition of the atmosphere, climate, and living organisms. The efficacy of the transport agents—water, ice, and air—is also dependent upon climate and on the tectonic conditions that determine the slopes existing on land surfaces. Whether sediments are deposited on the flooded margins of the continents or in the deep ocean basins is a function of the volume of material brought by the transport agent, the tectonic situation, and whether sea level is rising, falling, or stable. The kinds of material that accumulate in a particular place at a particular time depend on the sediment source and supply system, the tectonic setting, oceanographic conditions, and the efficacy of living organisms at extracting solutes. So many of these factors are interrelated through feedback mechanisms that a change in any one perturbs the entire system.

There are only two energy sources affecting the surface layers of the Earth: (1) radiation from the sun, which varies with orbital characteristics and may vary intrinsically, and (2) the interior heat of the Earth with its associated mantle motions driving the surface plates. Except for the relatively short-term orbital changes—precession (26,000 yr), obliquity (40,000 yr), and eccentricity (90,000 yr)—little is known of astronomical factors which might have altered the radiation reaching the Earth, thus affecting climate. Even less is known about the time scales of processes in the mantle. Mountain building may be episodic and seems to reach a climax every few hundred million years, but it is not clear that this is due to anything more than chance.

In the following we present a synopsis of the present state of knowledge of the volumes, masses, and compositions of major sedimentary reservoirs, present the global stratigraphic record of sedimentation in the ocean as it is known today, and describe the observed record of sea level changes. Then we discuss simple models which are suggested by modern theories and observations and determine the extent to which they can explain the observations or make predictions suitable for testing by other means.

2. Estimations of Sediment Volumes and Masses

In his classic work *Marine Geology*, Kuenen (1950) attempted to estimate the thickness of oceanic deposits, using three methods. For the first method, he used $\frac{1}{2}$ to $\frac{3}{4}$ cm/1000 yrs as an average sedimentation rate. He then determined that this was equal to $\frac{1}{6}$ cm/1000 yr solid. Multiplying by what was then thought to be the age of the ocean basins, 2×10^9 yr, he arrived at a total thickness of 3 km solid and a volume of 900×10^6 km^3. The second method relied on weathering processes producing sediment from igneous rocks. Citing Clarke's calculation of the proportions of shale, sand, and limestone expected to be produced from weathering of igneous rock, 20:3:2, he noted the very different ratio observed in fossil sediments, 20:12:14. Kuenen determined that if no lime had been deposited in the ocean, about three fourths of the shale and sand would have been lost off the continental blocks into the ocean. He cited the average thickness of Post-Precambrian sediment on North America as 0.8 km and estimated the total average thickness of sediment on the continental blocks, including Precambrian, at 1.2 km solid. The volume of sediment on the continents would then be 200×10^6 km^3. As a first approximation, he reasoned, the volume of oceanic sediments should be three times as large, 600×10^6 km^3. However, because lime has been lost to the ocean, Kuenen estimated the volume of oceanic sediment as 850×10^6 km^3 by this method. His third method was based on volcanic activity and denudation. Expanding arguments of Moss (1936), he assumed that about one third of the area of the continents has supplied the net sediments of the remainder of the continental cratons and margins, with 2 km of denudation of these Precambrian areas required for the production of the 650 m (solid) younger sediments on the continental cratons and an additional $2\frac{1}{2}$ km of denudation to produce the sediments of the continental margins. This is equal to a yearly amount of $\frac{1}{6}$ km^3 solid sediment. To this must be added the volume of volcanic material added, which had been estimated for the period since 1500 AD as 1 km^3/yr. Kuenen recognized that this figure was probably too high and used $\frac{1}{2}$ km solid as an annual production rate for volcanic material. From these annual increments of net sediment, he estimated the volume of deep-sea sediment deposited since the Precambrian at

250×10^6 km^3 and the Precambrian volume as 1000×10^6 km^3, for a total of 1250×10^6 km^3 oceanic sediment. He estimated the area of the continental slopes as 10% that of the Earth and the average thickness as 1.2 km solid, arriving at a volume of sediment on the continental slopes of 250×10^6 km^3 solid.

Kuenen's estimates involved the assumption of an ancient age for the ocean basins, and his estimates of sediment thickness were not borne out by subsequent seismic investigations. Had he used a modern estimate of 100×10^6 yr for the average age of the ocean basins, he would have proposed sediment volumes closely approximating those suggested by other investigators in recent years.

Most subsequent estimates of sediment volumes, masses, and compositions have relied on recognizing discrete subsets of the total sediment body and estimating their areal extent, thickness, and average porosity. The problem is so complex that only a few major sediment reservoirs are recognized and few authors have agreed in their numerical estimates. From the point of view of the dynamics of plate tectonics, the most important are:

1. Cratonic sediments, largely of Precambrian and Paleozoic age.
2. Sediments of Precambrian–Paleozoic geosynclines and Mesozoic–Cenozoic mobile belts.
3. Sediments of Mesozoic–Cenozoic trailing passive margin shelves.
4. Sediments of Mesozoic–Cenozoic continental slopes rises.
5. Pelagic sediments of the Mesozoic–Cenozoic ocean basins.

A. *Cratonic Sediments*

Poldervaart (1955), citing Kay (1951), estimated the area of the continental shield areas to be 105×10^6 km^2 or more. Kay had not given an actual areal measure for the cratonic continental area but estimated the area of eugeosynclines to be 25% of North America, of miogeosynclines to be 15% of North America, and the remaining area to be cratonic. It would appear that Poldervaart used the total area of continental land plus shelves (175×10^6 km^3) multiplied by 60% to arrive at the area of 105×10^6 km^2 for the shield areas. Poldervaart cited Kay's figure for the average thickness of sediments in such areas as 0.5 km solid and arrived at a volume of sediment of 52.5×10^6 km^3. The mass was estimated by Poldervaart as 140×10^{21} g. Kay (1951) had not in fact developed an independent estimate of the average thickness of sediment on continents but had converted Kuenen's (1950) figure of 0.8 km (solid) for Paleozoic and younger sediments of North America to 0.5 mile (solid). Correcting Poldervaart's citation of 0.5 km to 0.8 km to obtain a world value, the total volume of sediment on the cratonic areas would be 84×10^6 km. However Kuenen (1950) had given four different values for the average thickness of sediment on the continents: 0.8 km (solid) for the average thickness for North American Paleozoic and younger sediments (his page 387); 1.2 km (solid) for all sediments of North America, including Precambrian (his page 387); 650 m (solid) (page 387); and 1.5 km (page 396).

The first careful estimates were made by Ronov and Yaroshevsky (1969). Their areal measures for the shield areas and platform areas of the continents were determined by measuring the tectonic map of the world compiled by Tugolesov and Udintsev (1964) and were given as 29.4×10^6 and 66.9×10^6 km^2, respectively, totaling 96.3×10^6 km^2. In their Table 2 the average thickness of the sediments was given as 1.8 km. The volume was then calculated to be 135×10^6 km^3, which indicates an area of only 75×10^6 km^2 and they gave the mass as 0.35×10^{24} g, which involves assumption of porosity of only 5%.

The estimates of Ronov and Yaroshevsky have been refined in a recent paper (1977), taking into account data from larger-scale tectonic maps of the different continents. The new value was 35×10^6 km^2 for the shield areas and 55×10^6 km^2 for the platform areas; the value for the sediment volume was 210×10^6 km^3, including lavas amounting to 4.5% of the volume. Because the shield areas are essentially free of sediment, it would appear that the platform areas were assumed to have an average sediment thickness of 3.8 km. The average thickness over both shield and platform areas was given as 2.4 km, but according to our calculations from their data the value should be 2.33 km. The mass was given as 0.51×10^{24} g, so that the density was assumed to be 2.43 g/cm^3.

Gilluly and others (1970) calculated the volumes of sediments in and adjacent to the conterminous United States. Their estimate of the average thickness of sediments over the craton in the United States is 2.35 km. Using the areal estimates of Ronov and Yaroshevsky (1977) and the average of Gilluly and others (1970), the volume of sediment in platform areas for the world would be 129×10^6 km^3. The average of the two thickness estimates is close to 3 km, so that a volume of 165×10^6 km^3 is a suitable compromise. Assuming an average porosity of 20% and an average density for the solid phase of 2.7, the mass is 356×10^{21} g.

B. Geosynclinal Regions

Poldervaart (1955) estimated the area of the young folded belts at 42×10^6 km^2. He cited Kay's (1951) estimates of average thicknesses of sediments of 5 miles in eugeosynclinal regions and 2½ miles in miogeosynclinal belts. Poldervaart adopted 5 km solid (5½ km including pore space) as an average thickness of sediments and volcanics. He assumed that only 60% of the uppermost 5 km was really sediment, the rest being crystalline rocks, arriving at a volume of 126×10^6 km^3 solid (excluding pore space) and a mass of 340×10^{21} g.

Ronov and Yaroshevsky (1969) estimated the area of late Precambrian (Riphean) Paleozoic geosynclines as 24.4×10^6 km^2 and of "Mesocenozoic" geosynclines as 28.3×10^6 km^2 by measuring the world tectonic map as cited above. Although they mentioned 10 km as the average thickness of the sediments and volcanics, they gave the volume as 365×10^6 km^3. They estimated that 25.3% of the rocks are volcanic, so that the sediments would total 273×10^6 km^3. They gave the mass as 0.94×10^{24} g, having used 2.6 as the density.

Gilluly and others (1970) estimated the thickness of Phanerozoic rocks in the Cordilleran area of the United States at 6.24 km and in the Pacific area at 11.4 km. The areally corrected average thickness for these two regions combined is 8.4 km. Using Ronov and Yaroshevsky's estimate for the area and Gilluly and others' thicknesses of 11.4 and 8.4 km, the volumes would be 416×10^6 and 307×10^6 km^3, respectively.

Ronov and Yaroshevsky (1977) revised their estimates but still assigned an area of 28.3×10^6 km^2 to Mesozoic–Cenozoic Geosynclines exposed on land. However they gave an area of 30.7×10^6 km^2 to Riphean–Paleozoic Geosynclines exposed on land. They gave the volumes as approximately 260 × 10^6 km^3 for Mesozoic–Cenozoic geosynclines (average thickness 9.2 km) and 270 × 10^6 km^3 for Riphean–Paleozoic geosynclines (average thickness 8.8 km). The masses are 657 × 10^{21} and 683 × 10^{21} g, respectively, using a density of 2.53. They estimated lavas to be 21.9% by volume of the total, so that the sediment volume becomes 203 × 10^6 km^3 for the Mesozoic–Cenozoic geosynclines; the sediment masses become 503 × 10^{21} and 524 × 10^{21} g, respectively.

Ronov and Yaroshevsky (1977) also assigned areas to the Mesozoic–Cenozoic (24.0 × 10^6 km^2) and Riphean–Paleozoic (14.9 × 10^6 km^2) geosynclines presently flooded by shelf and slope seas. These total a larger area than that they ascribed to cratonic platforms presently flooded (26.0 × 10^6 km^2).

C. *Continental Shelves and Coastal Plains*

Poldervaart (1955) cited 30 × 10^6 km^2 as the area of continental shelves and assumed an average of 4 km solid thickness, giving a volume of 120 × 10^6 km^3 and a mass of 324 × 10^{21} g.

Menard and Smith (1966) gave the area of the ocean less than 200 m deep as 27.1 × 10^6 km^2, most of which is continental shelf. They give the area of the continental shelf and slope together as 55.4 × 10^6 km^2 and the continental rise as 19.2 × 10^6 km^2, for a total of shelf–slope–rise area of 74.6 × 10^6 km^2. The "subcontinental region" of Ronov and Yaroshevsky (1969) was estimated at 65.5 × 10^6 km^2 and was reevaluated by them in 1977 as 64.9 × 10^6 km^2. Neither of these values corresponds to the shield and slope of Menard and Smith and must include the upper part of what is termed continental rise in the physiographic terminology of Menard and Smith.

The shelf areas along the oceans have been planimetered and values for 10° latitudinal increments given by Hay and Southam (1977). The trailing passive margin shelves of the oceans are 21.2 × 10^6 km^2, with 10.8 × 10^6 km^2 in the Atlantic and its tributary seas, 6.3 × 10^6 km^2 in the Indian Ocean, and 4.1 × 10^6 km^2 in the Pacific and its tributary seas. The trench-bordered shelves have an area of 6.3 × 10^6 km^2 and are almost entirely in the Pacific.

The width of coastal plains on land is about half that of the adjacent trailing passive margin shelves, so that the area of coastal plain and shelf sedimentation produced as a result of the breakup of Pangaea is about 31.8 × 10^6 km^2.

Determination of the average thickness of the sediments of the shelves is difficult at present because the nature of the deeper layers is in question. We arrived at an average maximum thickness of 3.7 km by averaging the maximum shelf sediment thickness indicated on profiles along the eastern margin of North America presented by Rabinowitz (1974). Extrapolating to a world figure by assuming a wedge of coastal plain and shelf sediment of triangular section, we estimate the volume of sediments which have accumulated on the Mesozoic–Cenozoic trailing margins as 59 × 10^6 km^3, or about 47 × 10^6 km^3 solid. This is less than half of Poldervaart's estimate of 120 × 10^6 km^3 solid, which would appear to be undoubtedly too high because it would require an average maximum thickness of 10 km. Using the map of sediment thickness of Emery and Uchupi (1972, their Fig. 188) an average maximum thickness of 6 km sediment for the Atlantic and Gulf Coast shelves seems more likely. Extrapolated to a global shelf sediment volume estimate, this yields a volume of 95×10^6 km^3, or 76×10^6 km^3 solid assuming a porosity of about 20%, as would be indicated by the studies of Atwater and Miller (1965). This estimate is considered the most likely for the global value based on present knowledge and corresponds to a mass of 205×10^{21} g.

D. Continental Slope and Rise

The slope and rise mark the transition between the continent and ocean basin. Menard and Smith (1966) joined the shelf and slope as a single physiographic province. Others have defined the boundaries at a variety of places using a variety of criteria. The slope and rise are the least-known sediment accumulations, and estimates of the sediment volumes must still be regarded as largely speculative.

As noted above, Kuenen (1950) estimated the volume of slope and rise sediment to be 250×10^6 km^3 solid. It is not clear exacly how he arrived at this figure.

Poldervaart (1955) cited the area covered by hemipelagic sediments as 63×10^6 km^2 and the average thickness as 5 km including pore space, referring to Kuenen (1950, page 396) and Kay (1951, page 92). Kuenen's figure was actually his estimate of maximum thickness; Kay offered no estimate. Poldervaart assumed that the 315×10^6 km^3 of hemipelagic sediment contained 20% pore space, yielding an estimate of 252×10^6 km^3 solid and a mass of 680×10^{21} g.

Southam and Hay (1977) estimated the volume of sediment in the continental slope and rise by assuming them to be underlain by cool oceanic crust and by assuming that each elevation increment above a base level defined by Sclater and Francheteau's (1970) curve to be due to sediment loading and to be isostatically adjusted. They used the data of Menard and Smith (1966) on areas of slope and rise between depth increments and assigned to the Pacific rises an area of 7.5×10^6 km^2, to the Indian Ocean rises an area of 9×10^6 km^2, and to the Atlantic rises an area of 11×10^6 km^2. Assuming an average porosity of 15% or average density of 2.5, they determined a volume of 248×10^6 km^3 including pore space, which equals 211×10^6 km^3 solid. Recalculated, assuming pore space of 20%, the estimate becomes 220×10^6 km^3 because of the isostatic adjustment, or 176×10^6 km^3 solid or 475×10^{21} g.

Ronov and Yaroshevsky (1969) gave the area of shelf and slope sediments as 65.5×10^6 km^2, with an average thickness of 2.9 km, yielding an estimated volume of 190×10^6 km^3 and a mass of 0.48×10^{24} g. In their 1977 paper they revised the area to 64.9×10^6 km^2 and the estimated sediment volume to 160×10^6 km^3, indicating an average thickness of only 2.5 km. The mass was calculated to be 0.40×10^{24} g.

Gilluly and others (1970) estimated the sediment offshore the eastern United States to be about 10×10^6 km^3 and in the northern half of the Gulf of Mexico to be 12×10^6 km^3. Emery and Uchupi (1972) estimated the offshore sediment volume in these areas to be 41×10^6 km^3. An extrapolation to obtain a global figure from these figures yields sediment volume values of 412×10^6 and 769×10^6 km^3, respectively. These values are far higher than any other estimates, probably because they include the anomalous Gulf of Mexico. Using the estimate of Gilluly and others for the Atlantic coast and extrapolating, a value of 187.5×10^6 km^3 (150×10^6 km^3 solid) is obtained, but this is probably too low because it excludes any of the effects of marginal seas. Estimates based on Emery and Uchupi's (1972) data suggest a global volume of 375×10^6 km^3, or 300×10^6 km^3 solid.

Obviously the volume of slope and rise deposits is the least well known of the sediment reservoirs. Ronov and Yaroshevsky's estimates of 1969 and 1977 seem to us to be too low, and estimates based on an extrapolation of Emery and Uchupi's data seem too high. A compromise estimate of about 280 × 10^6 km^3, or 255 × 10^6 km^3 solid, seems reasonable in the present state of knowledge. This corresponds to a mass of 607 × 10^{21} g.

E. Pelagic Sediments

Kuenen's (1950) estimates of the volume of pelagic sediments have been discussed above. They were based on arguments involving the assumption of a Precambrian age for the ocean basins and are much higher than later estimates based on seismic determinations of sediment thicknesses.

Poldervaart (1955) used 268 × 10^6 km^2 as the area covered by pelagic sediments and 0.6 km as their average thickness including pore space. He believed that pelagic sediments with 70% pore space at the surface and an assumed 35% pore space at 600 m depth would have an average porosity of 50%, so that the thickness of the solids would be 0.3 km, giving a volume of 80.4 × 10^6 km^3. Poldervaart noted that if one assumed a long-term sedimentation rate only half that of the present rate, the present sediment accumulation would be attained in only 200 million years. He also suggested that "the conspicuous development of *Globigerina* . . . some 150 million years ago" might be significant. Poldervaart estimated the mass of the pelagic sediments to be equivalent to 217 × 10^{21} g.

Ronov and Yaroshevsky (1969) used 300 × 10^6 km^2 as the area of pelagic sedimentation and 0.4 km as the average thickness, giving a volume of 120 × 10^6 km^3. They appear to have assumed a porosity of 40%, which would yield a volume of 72 × 10^6 km^3 solid and, multiplied by a density of 2.7, their estimated mass of 0.19 × 10^{24} g. In their latest paper (1977) they did not modify these estimates.

Southam and Hay (1977) gave the area over which pelagic sediments rain as 311 × 10^6 km^2 and assumed an average thickness of 0.5 km, leading to a volume of 155.5 × 10^6 km^3. They assumed an average porosity of 25% and calculated a mass of 358 × 10^{21} g. Excluding the area of the continental rises where the pelagic sediment is mixed with terrigenous sediment, the volume is reduced to 142 × 10^6 km^3 and the mass to 327 × 10^{21} g. The pelagic sediment which accumulates in the continental rises is about 13.5 × 10^6 km^3, or 31 × 10^{21} g.

An average thickness of 0.5 km seems reasonable in the present state of knowledge, but the estimate of porosity can be refined as a result of study of cores taken by the Deep Sea Drilling Project. Using 35% as an average porosity as suggested by the studies of Hamilton (1976), the volume of pelagic sediment, exclusive of the continental rises, becomes 92 × 10^6 km^3 solid and the mass becomes 249 × 10^{21} g.

F. Summary of Volumes and Masses of the Major Reservoirs

The ranges of estimates of volumes and masses for the major sedimentary reservoirs are indicated in Fig. 1; the values chosen for developing the subsequent discussions are circled in the figure. See also Table I.

The area of cratonic shields is taken to be 29.4 × 10^6 km^2, that of cratonic

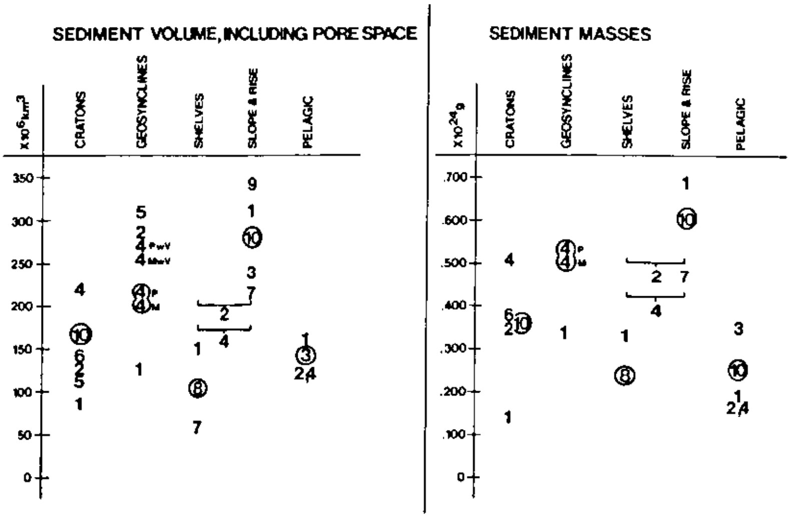

Fig. 1. Estimates of the volumes and masses of cratonic, geosynclinal, shelf, slope and rise, and pelagic sediments. (1) Poldervaart (1955); (2) Ronov and Yaroshevsky (1969); (3) Southam and Hay (1977); (4) Ronov and Yaroshevsky (1977); (5) extrapolated from thickness estimates of Gilluly and others (1970) and areal estimates of Ronov and Yaroshevsky (1969); (6) extrapolated from thickness estimates of Gilluly and others (1970) and areal estimates of Ronov and Yaroshevsky (1977); (7) Southam and Hay (1977) recalculated assuming 20% pore space; (8) extrapolated from thicknesses indicated by Emery and Uchupi (1972) and areas estimated by Southam and Hay (1977); (9) extrapolated from estimates of Emery and Uchupi (1972); (10) compromise estimate adopted. Volumes include pore space.

platforms as 66.9 × 10^6 km^2; volume of cratonic sediments is taken to be 165 × 10^6 km^3, and their mass is taken as 356 × 10^{21} g. The area of Precambrian and Paleozoic geosynclines on land is taken to be 30.7 × 10^6 km^2. The volume of sediments in them is accepted as 211 × 10^6 km^3, and their mass is taken as 524 × 10^{21} g. Although some minor amount of younger sediments may be included in the totals of 127 × 10^6 km^2, 376 × 10^6 km^3, and 880 × 10^{21} g, they are essentially the values for the amount of Precambrian and Paleozoic sediments preserved.

The area of Mesozoic–Cenozoic geosynclines on land is accepted as 28.3 × 10^6 km^2, the volume of sediments in them is accepted as 203 × 10^6 km^3, and their mass is 503 × 10^{21} g. The area of passive margin shelves formed during the Mesozoic–

TABLE I

Area, Volume, and Mass of Sediments in the Major Sedimentary Reservoirs

Sediments	Area ($\times 10^6$ km^2)	Volume ($\times 10^6$ km^3)	Mass ($\times 10^{21}$ gr)
Cratonic	96.3	165	356
Precambrian–Paleozoic geosynclines	30.7	211	524
Mesozoic-Cenozoic geosynclines	28.3	203	503
Passive margin	31.8	95	205
Active margin	6.3	19	41
Slope + rise	27.5	280	607
Pelagic	284	142	249

Cenozoic is approximately 31.8×10^6 km^2, their volume is 95×10^6 km^3, and their mass is about 205×10^{21} g. The area of active margin at trench-bordered shelves is about 6.3×10^6 km^2, their volume about 19×10^6 km^3, and their mass about 41×10^{21} g. The area of the continental slopes and rises is taken to be 27.5×10^6 km^2, their volume is estimated at 280×10^6 km^3, and their mass is 607×10^{21} g. The area covered by pelagic sediments is 284×10^6 km^2, with a volume estimated to be 142×10^6 km^3, and their mass is 249×10^{21} g. These Mesozoic–Cenozoic reservoirs total 739×10^6 km^3, or 1605×10^{21} g.

Because the sediments reservoirs are essentially Mesozoic–Cenozoic in age, almost two thirds of the total sediment present on the earth (2485×10^{21} g) has been sedimented onto the continental margins and deep sea since the Paleozoic. Because there are no appreciable sediment sources within the ocean basins themselves, the total of 1605×10^{21} g has been removed from preexisting continental areas through either reworking of older sediments or weathering of igneous rocks. The average rate of sediment delivery to the sea would be 70×10^{14} g/yr, compared with the present rate of 249×10^{14} g/yr cited by Garrels and Mackenzie (1971).

3. Composition of the Sedimentary Reservoirs

Poldervaart (1955) discussed average compositions for each of the sedimentary reservoirs he estimated. The average compositions were determined from a number of sources, combining data where appropriate.

Ronov and Yaroshevsky (1969) presented more detailed average compositions based on much larger numbers of sample analyses as well as on calculations of relative volumes of materials on the continents determined by measuring paleogeographic–lithologic maps for the Soviet Union. Their data were revised in 1977.

Sibley and Wilband (1977) calculated compositions for continental platform, geosyncline, "subcontinental," and several pelagic sediments, using data from Ronov and Yaroshevsky (1969) but with differing proportions for the different sediment types. The first three columns of Table II are the average compositions of cratonic sediments given by Poldervaart (1955), Ronov and Yaroshevsky (1977) (as recalculated on a water-free basis), and Sibley and Wilband (1977). There are significant differences between these, introduced largely because Ronov and Yaroshevsky estimated about 25% more carbonate rock (limestone and dolomite) than Poldervaart. Columns 4 through 6 of Table II give the average compositions of geosynclinal sediments of Poldervaart (1955), Ronov and Yaroshevsky (1977), and Sibley and Wilband (1977). The estimates are remarkably similar, in much closer agreement than those for cratonic sediments, differing chiefly in the lower proportion of carbonate in the Sibley and Wilband composition.

Columns 7 through 10 of Table II give the average compositions of pelagic sediments as determined by Poldervaart (1955), El Wakeel and Riley (1961), Ronov and Yaroshevsky (1977), and Southam and Hay (1977). Sibley and Wilband (1977) calculated compositions for pelagic sediments based on the compositions of El Wakeel and Riley (1961) for mixtures containing different amounts of carbonate. El Wakeel and Riley (1961) calculated the composition of pelagic sediments by multiplying the average composition of calcareous ooze, siliceous ooze, red

TABLE II
Compositions of Sedimentary Reservoirs

	Continental Platforms			Geosynclines			Pelagic Sediments				Subcontinental (Shelf, Slope and Rise)	
	I Polder-vaart (1955)	II Ronov and Yaro-shevsky (1977)	III Sibley and Wilband (1977)	IV Polder-vaart (1955)	V Ronov and Yaro-shevsky (1977)	VI Sibley and Wilband (1977)	VII Polder-vaart (1955)	VIII El Wakeel and Riley (1961)	IX Ronov and Yaro-shevsky (1977)	X Southam and Hay/ Sujowski (1977)	XI Ronov and Yaro-shevsky (1977)	XII Sibley and Wilband (1977)
SiO_2	60.6	46.27	49.5	51.9	50.41	55.2	28.5	42.7	42.59	24.1	46.27	53.3
TiO	0.4	0.65		0.5	0.72		0.4		0.65	0.3	0.65	
Al_2O_3	8.9	10.10	10.7	11.4	13.28	12.0	8.1	12.3	11.85	7.4	10.10	11.6
Fe_2O_3	2.4	2.62		2.6	2.59		5.0		4.84	6.3	2.62	
Fe			3.2			3.6		4.1				3.5
FeO	1.2	2.11		2.0	3.16		—		1.02	—	2.11	
MnO	tr.	0.07		tr.	0.11		0.6		0.36	0.7	0.07	
MgO	2.9	3.94	3.3	3.8	3.21	3.2	1.8	2.9	3.10	1.3	3.94	3.2
CaO	10.6	14.57	12.3	12.6	13.11	8.3	30.5	14.5	17.51	40.2	14.57	9.6
Na_2O	0.8	1.32	0.9	1.3	1.70	1.0	0.8	1.1	1.19	1.4	1.32	1.0
K_2O	2.1	2.29	2.2	2.4	1.98	2.5	1.2	2.1	2.13	1.1	2.29	2.4
P_2O_5	0.1	0.11		0.1	0.16		0.2		0.16	0.3	0.11	
Corg		0.61			0.49		—		0.37	—	0.61	
CO_2	10.0	12.38	10.8	11.4	8.95	7.5	22.9	11.7	13.92	16.8	12.38	8.6
SO_2		1.73			0.91				—		1.73	
S		0.29			0.15				0.03	0.29		
Cl		0.85			0.12				—		0.85	
F		0.06			0.04				0.05		0.06	

clay, and volcanogenic sediments by the proportion of the area of the world ocean basins which they occupy. Ronov and Yaroshevsky (1969, 1977) used the same method for calculating the average compositions of pelagic sediment; but for cratonic and geosynclinal sediments, they based the calculation on the relative volume proportions of the different sediments. Poldervaart (1955) used the average compositions of red clay, siliceous ooze, and calcareous ooze recalculated from unpublished determinations of Rubey (1951). He assigned to them relative sedimentation rates of 1 (red clay), 1.25 (siliceous ooze), and 3 (calcareous ooze) and multiplied these values by the areas occupied by each (102.2×10^6, 38×10^6, and 127.9×10^6 km^2, respectively) to determine the relative volumes as 19.2, 8.9, and 71.9% of the total.

Because El Wakeel and Riley (1961) and Ronov and Yaroshevsky (1969, 1977) used area rather than volume to calculate the average composition of oceanic sediment, because the differential sedimentation rates of terrigenous, calcareous, siliceous, red clay, and volcanic components at the present time are well known, and because the areal distribution is known to be a function of physical oceanographic conditions, we do not consider their estimates of the average composition to be valid.

The largest unknown is the average composition of sediments that have accumulated on passive margin shelves, slopes, and rises since the breakup of Pangaea. Ronov and Yaroshevsky (1969, 1977) assumed that the composition of the sediments in the "subcontinental" regions (continental shelves, slopes, and rises) would be the same as that of sediments on the continents. Sibley and Wilband (1977) calculated a composition for "subcontinental" sediment, assuming it to be a mixture of two thirds geosynclinal sediment and one third continental sediment, arriving at the composition shown in column 12 of Table II. These estimates for "subcontinental" sediment composition are regarded as highly speculative. Deep seismic exploration suggests that salt deposition may be a characteristic of the early opening oceans. The Red Sea, early North Atlantic, and early South Atlantic are thought to be extensively underlain by evaporites, and this volume could be very significant, possibly amounting to as much as 5 to 10% of the total volume of sediment in these areas (Kinsman, 1974). This is particularly important because Ronov and Yaroshevsky (1969) had estimated the abundance of evaporites at 2% of the sediments on continents and (in 1977) had revised this to 2.8%. For geosynclinal areas, they cited (1969, 1977) the proportion of evaporites as 0.3%. Using their estimates of volumes, these values would indicate 5.9×10^6 km^3 of evaporites on the continental platforms and 1.6×10^6 km^3 in geosynclinal regions. Because they considered the proportions of rocks in the subcontinental regions to be the same as those exposed on land, a volume of evaporites in the shelves, slopes, and rises of 1.6×10^6 km^2 is suggested. However judging from the maps and cross sections of the Gulf of Mexico by Antoine et al. (1974) and Ballard and Feden (1970), there may be as much as 4×10^6 km^3 of evaporites underlying the Gulf of Mexico and its immediate margins. Similarly the Mediterranean may contain 1×10^6 km^3 of evaporites and the Red Sea 0.5×10^6 km^3 or more. The quantities in the Atlantic are much more difficult to estimate because not all piercement structures indicate the presence of salt (Lancelot and Embley, 1977). The sizes of the North Atlantic and South Atlantic basins and the thicknesses of evaporites reported from better-known sections by Evans (1978) however indicate that there may be up to 5×10^6 km^3 in the Atlantic margins, 4×10^6 km^3 in the South Atlantic, and 1×10^6 km^3 in the North Atlantic. This increment of more than 10×10^6 km^3, added to the world inventory by exploration in the past decade, more than doubles the known amounts of evaporites and has great significance for the salt balance of the ocean.

Estimates of the mass of sedimentary rock as a function of time have, of necessity, relied most heavily on information from the land areas exposed on the continents. Ronov (1959) estimated the mass of Devonian through Jurassic rocks by period, based on measurement of existing sedimentary volumes on the better-known parts of the Soviet Union. Gregor (1968) recalculated these estimates by taking into account the maximum thickness data presented by Holmes (1965). Garrels and Mackenzie (1971) based their estimates of sedimentary mass through time on the assumption that the mass would be proportional to the maximum thickness of strata as given by Kay (1955). All of these estimates assumed that the information from land areas could be extrapolated to the sedimentary mass of the Earth as a whole, but this is obviously untrue in the context of plate tectonics, which requires that about two thirds of the total sedimentary mass on Earth has been deposited since the beginning of the Mesozoic. Ronov (1964) published a very interesting diagram showing the volume percent of different types of sedimentary rocks as a function of time. This diagram emphasized the importance of jaspilites in the Precambrian, illustrated the very striking change in the relative proportions of dolomite and limestone from the Paleozoic to Mesozoic, and suggested that the proportion of all carbonate rocks, limestones, and dolomites has been sharply reduced since the Paleozoic. Reproduced in subsequent works (Ronov, 1968; Garrels and Mackenzie, 1971), this diagram has been the basis of much discussion on the evolution of sedimentary rocks. It is present here in modified form as Fig. 2, using Ronov and Yaroshevsky's (1977) estimates of the volumes of sedimentary reservoirs established above and Poldervaart's (1955) volume proportions of the pelagic sediments.

The large quantity of calcareous sediment which has been accumulating as pelagic ooze during the past 100 million years may well account for the decline in carbonate sediment since the Paleozoic suggested by Ronov (1964) on the basis of the record preserved on the continents.

The large quantity of evaporites in the continental margins undoubtedly includes large amounts of calcium sulfate, acting as a special sink for Ca. Interestingly, if our estimate of the volume of 10×10^6 km^3 evaporites in the 375×10^6 km^3 of the shelf slope and rise sediment is correct, the evaporites are 2.7% of the sedimentary volume. Ronov and Yaroshevsky (1977) cited the volume of evaporites in continental shield and platform areas as 2.8% of the sedimentary volume and speculated that the proportions in the shelf, slope, and rise would be the same.

4. The Stratigraphic Record of Sedimentation in the Ocean

Cores recovered by the Deep Sea Drilling Project have provided much information about deep-sea sediments. Because of the uniform manner in which samples from the Deep Sea Drilling Project have been handled, they now constitute the most homogeneous body of data on sedimentary rocks available. Unfortunately only two chemical determinations were routinely made on Deep Sea Drilling

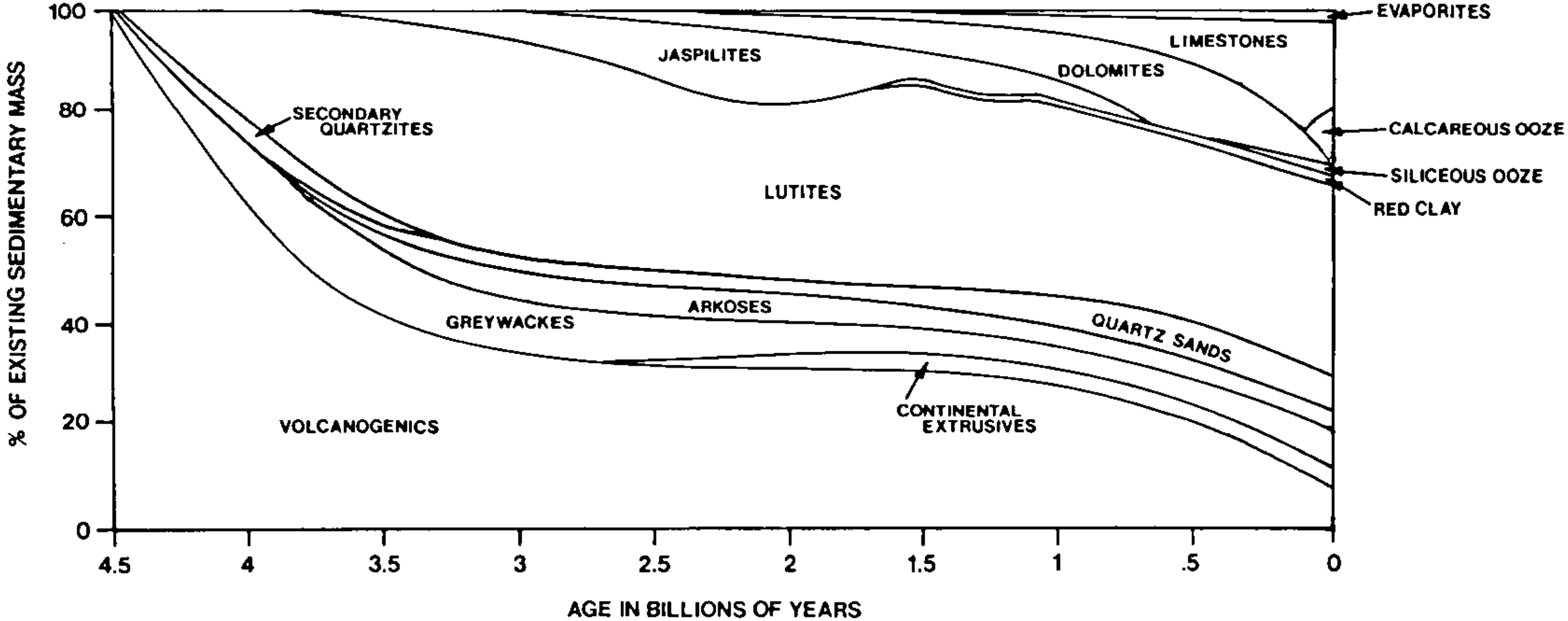

Fig. 2. Evolution of sedimentary rocks, modified after Ronov (1964). The vertical scale represents the proportion of sedimentary rock of a given type forming at a given time.

Project materials: the percentage of carbonate and the percentage of organic carbon.

Using raw data on sediment thickness and the percentage carbonate from as many drill sites as possible, Davies and others (1977) estimated sedimentation rates for the following intervals of Cenozoic time: Quaternary, Late Pliocene, Early Pliocene, Late Miocene, Middle Miocene, Late Oligocene, Early Oligocene, Late Eocene, Middle Eocene, Early Eocene, Late Paleocene, and Early Paleocene. They noted significant, synchronous differences in sedimentation rates in the Atlantic, Pacific, and Indian Oceans, with the Middle Eocene and Middle Miocene–Quaternary intervals being times of high sedimentation rate and the Paleocene–Early Eocene and Late Eocene–Early Miocene being times of low sedimentation rate. They estimated the difference between times of low and high sedimentation rate, to be a factor of about 5 and suggested that, because this difference occurs on time scales too long to be the result of sea level trends or changes in the mixing rate of the ocean, they must be the result of changes in the river inputs and ultimately reflect changes in the weathering rates of the continents.

The sedimentation rates are presented in Figs. 3, 4, and 5 for both the Cenozoic and the Cretaceous, to the extent the data will allow. Curves for the Atlantic, Pacific, and the world ocean are shown. Indian Ocean data are sparse for the Cretaceous, and no useful information could be added to the analysis of Davies and others (1977). The diagrams are semilog plots so that organic carbon can also be shown. These curves, like those presented by Davies and others (1977), are based on raw averages of all determinable thicknesses of sediments for the same intervals of the Cenozoic and for the stages of the Cretaceous, uncorrected for compaction. Sedimentation rates for the Cenozoic were recalculated using time lengths for the intervals corresponding to the Berggren (1972) time scale. For the Cretaceous, the Van Hinte (1976) time scale was used. These figures also show noncarbonate sedimentation as a distinct component.

Figure 3 shows the record for the Atlantic Ocean for sedimentation during the past 110 million years. The trend for total sedimentation differs slightly from that published by Davies and others (1977) because the Middle Miocene was considered to be longer and the Early Miocene to be shorter, with the result that sedimentation rates appear to rise more evenly during the Early Neogene. Presentation in semilog format allows the difference in sedimentation rates between the Paleocene and the Oligocene to be more clearly displayed. Early Paleocene sedimentation rates are only a tenth of the Quaternary rate. Although the curve is not corrected for compaction, this effect can be roughly estimated from the generalized data on porosity-depth relationships presented by Hamilton (1976), with the result that the values for the Pleistocene should be reduced by about 30% and those for the Pliocene by about 20%, to bring them into accord with the 35 to 40% pore space included in the older values.

The noncarbonate component, which includes terrigenous sediment, red clay, and siliceous ooze, shows striking trends. It does not rise and fall as a fixed proportion of the total sediment but constitutes a larger part of the total during times of high sedimentation rate and a lesser part of the total during times of lower sedimentation rate. The drill sites from which these data were compiled included midocean ridge flanks, abyssal plains, and outer continental rise areas, so that the composite sedimentation rate is not one for pelagic sediments exclusively. The

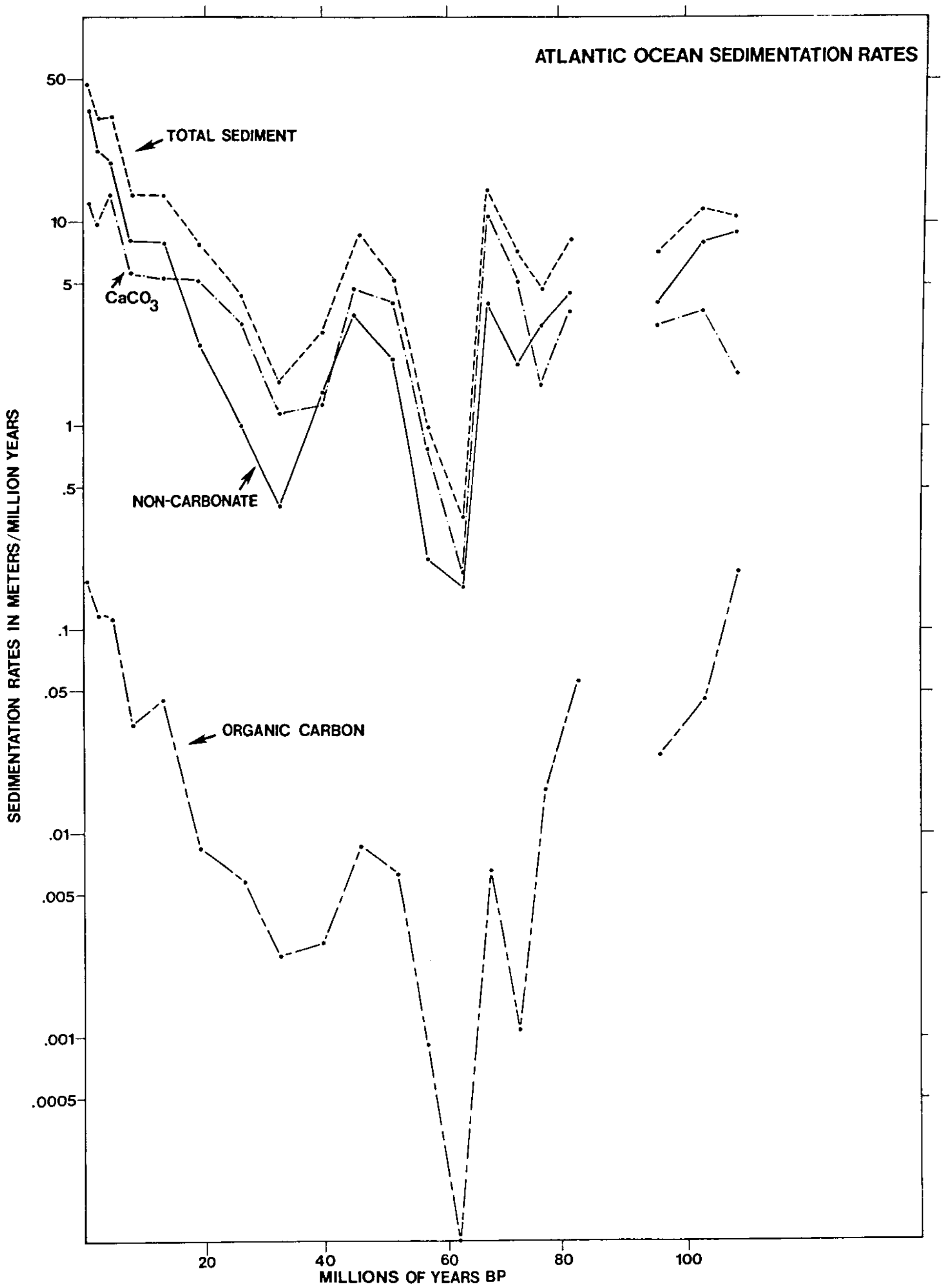

Fig. 3. Average sedimentation rates for the Atlantic Ocean estimated from the data of the Deep Sea Drilling Project, uncorrected for compaction.

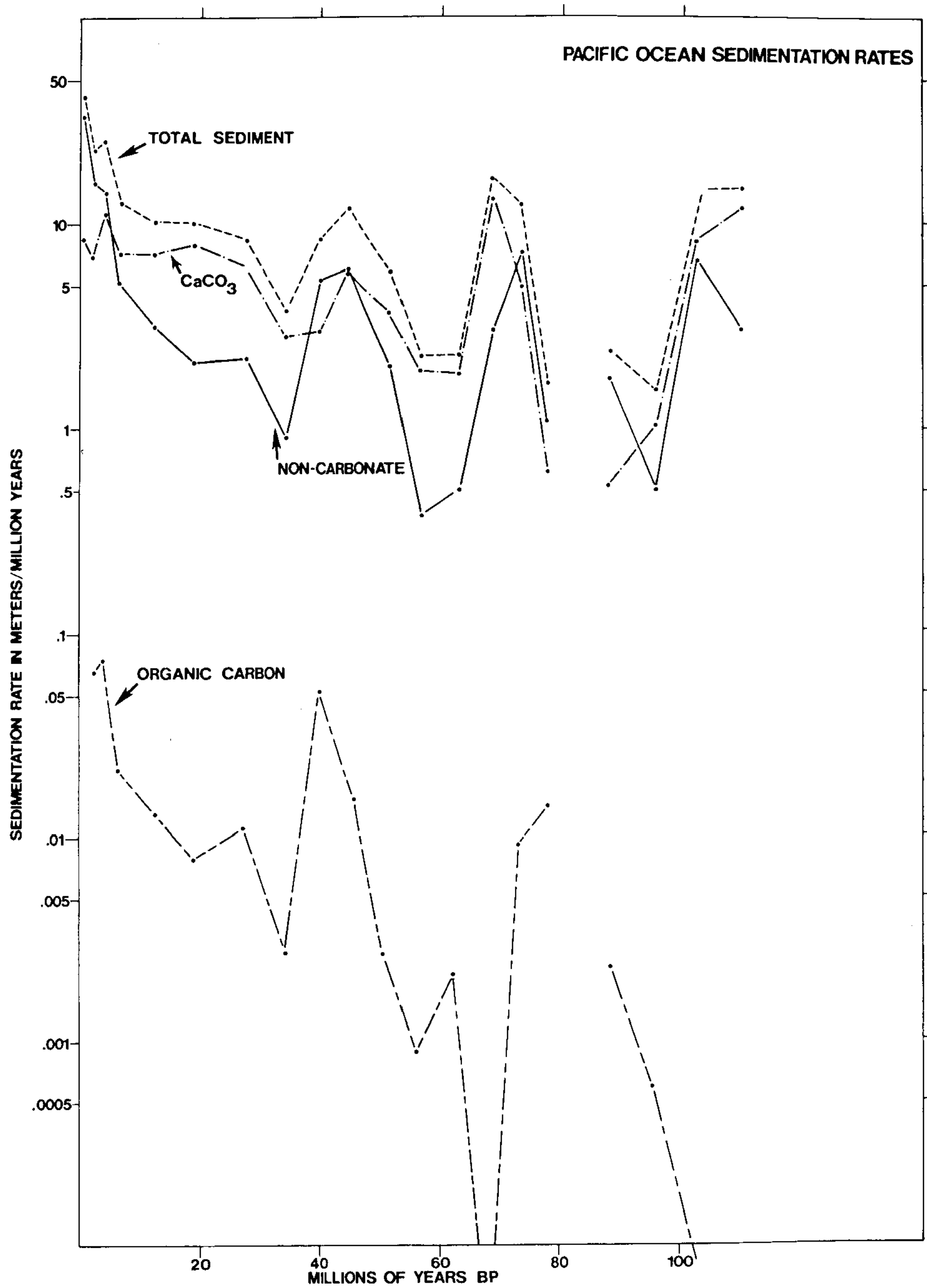

Fig. 4. Average sedimentation rates for the Pacific Ocean estimated from the data of the Deep Sea Drilling Project, uncorrected for compaction.

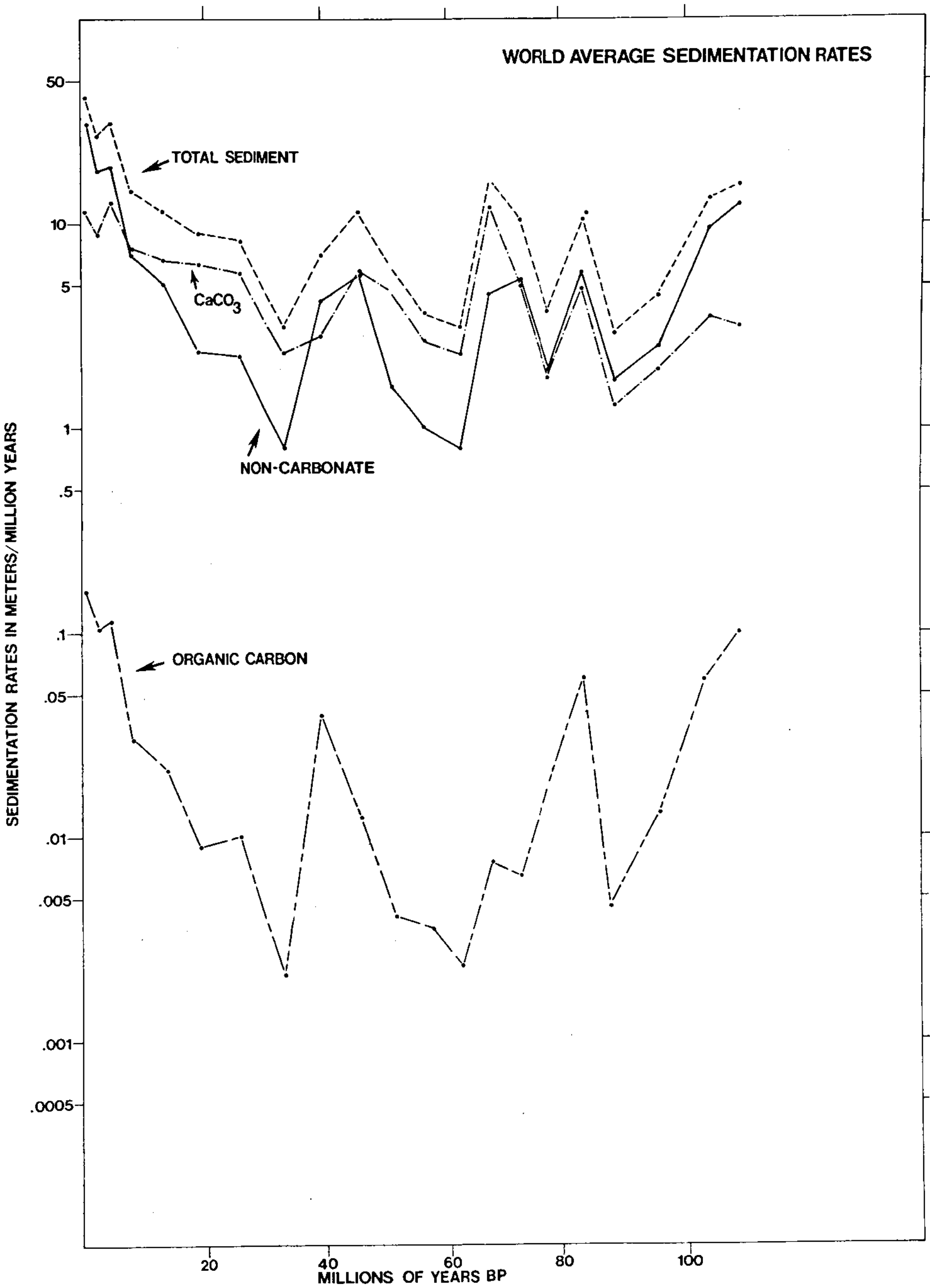

Fig. 5. Average sedimentation rates for the world ocean estimated from the data of the Deep Sea Drilling Project, uncorrected for compaction.

high proportion of noncarbonate sediment in the Middle Miocene–Quaternary interval, exceeding the carbonate, suggests that supply conditions to the Atlantic may have been different during this time from those during the earlier Cenozoic and Late Cretaceous, when the carbonate fraction is almost always greater than the noncarbonate fraction. Only during the Cretaceous prior to the Campanian and during the Late Eocene did the supply of noncarbonate exceed that of carbonate.

The trend of the carbonate sedimentation rate curve follows that of the total sedimentation rate. Particularly striking is the sedimentation rate high of about 70% carbonate in the Maastrichtian, the highest carbonate sedimentation rate average for the Atlantic if the relative lack of compaction of Pliocene and Quaternary sediments is taken into account. Fluctuations of the carbonate curve are more damped than those of the noncarbonate curve, and the lowest rate is only about one fifth of the highest one.

The organic carbon sedimentation rates follow the trends of the noncarbonate sedimentation rate curve. Note particularly the decrease in both noncarbonate sediment and organic carbon in the Campanian, although the carbonate and total sedimentation rates for the Campanian are higher than for the Santonian. The organic carbon sedimentation rate fluctuates much more strongly than the other known fractions, with the minimum rate in the Early Paleocene being only about a thousandth of that of the Quaternary.

Figure 4 shows the same curves for the Pacific Ocean. The trends tend to be much the same but are somewhat damped, as would be expected from the greater size of the ocean. Noncalcareous sedimentation exceeds calcareous sedimentation in the Campanian–Santonian. Middle and Late Eocene and the Pliocene and Quaternary show essentially the same pattern as in the Atlantic. The Maastrichtian carbonate sedimentation rate is the highest of the period for which data are available, while the Paleocene carbonate sedimentation is less than the Oligocene, as in the Atlantic. The organic carbon sedimentation rate trend generally follows that of noncarbonate sedimentation but is lower than that of the Atlantic. It is interesting to note that the Maastrichtian sediments of the Pacific are virtually devoid of organic carbon, an exception to the general rule that organic carbon sedimentation varies as a function of the noncarbonate sedimentation rate.

Figure 5 shows world average sedimentation rates and includes information from the Indian Ocean. The general global trends emerge: (1) relatively high sedimentation rates in the Aptian–Albian, Coniacian, Campanian–Maastrichtian, Middle Eocene, and Late Miocene–Quaternary; (2) relatively low sedimenation rates in the Coniacian–Turonian, Santonian, Paleocene–Early Eocene, and Late Eocene–Early Oligocene; (3) intermediate sedimentation rates in the Late Oligocene–Early Miocene; (4) noncalcareous sedimentation exceeding the calcareous one in the Campanian and older Cretaceous, the Late Eocene, and in the Pliocene–Quaternary; (5) sedimentation overall fluctuating by a factor of 10, noncarbonate by a factor of 15, carbonate by a factor of 5, and organic carbon by a factor of 100.

5. The Stratigraphic Record from the Continental Margins

At present, knowledge of the continental margins, except for the shelf, is derived chiefly from geophysical investigation. Relatively few data are available from drilling to determine the ages and lithology and calibrate the geophysical data. Hence knowledge of the sedimentary rocks of the continental slopes and rises is very rudimentary at present. Much promise lies in the work of Vail and others (1977), which suggests that calibrated multichannel seismic sections can permit detailed interpretations of the stratigraphy over large areas. Since the early pioneering work of Drake et al. (1959), Ewing et al. (1966), and Sheridan et al. (1969, 1970), a large number of seismic studies of continental margins have been published. In recent years these include works by Emery and Uchupi (1972), Amoco Canada and Imperial Oil (1974), Burk and Drake (1974), Beck and Lehner (1974), Lehner and de Ruiter (1977), Keen (1974), Keen and Keen (1974), Scholl and Marlow (1974), Mattick et al. (1974, 1978), Case (1974a, 1974b), Sheridan (1974, 1976), Emery et al. (1975), Martin and Case (1975), Schlee et al. (1976), Thompson (1976), Uchupi et al. (1976, 1977), and many others, including long profiles published by the U.S. Geological Survey and the AAPG. Unfortunately almost none has a sufficiently detailed stratigraphic interpretation to permit comparison of depositional rates in the margin with those indicated by an analysis of the results of the Deep Sea Drilling Project. The stratigraphic analysis of Gulf of Mexico and offshore West African sequences by Todd and Mitchum (1977) has adequate resolution but is restricted to Triassic–Early Cretaceous strata and does not overlap the interval for which global sedimentation rates are known from the Deep Sea Drilling Project results. Seismic profiles show structures interpreted as reefs and suggest the presence of evaporites in the margins of the Atlantic basins as noted above. Until drilling will have proved the nature of these materials, discussion of the sedimentary rocks of the slope and rise must remain highly speculative. It is already evident nevertheless that the general principle of sediment accumulation rates decreasing with time after the breaking apart of a passive margin is confirmed by a large number of geophysical sections.

6. The Observed Record of Sea Level Changes

The record of sea level through geologic time has been reconstructed in two ways: (1) by analysis of transgressive–regressive sequences and (2) by measuring the continental surface flooded at a given time using paleogeographic maps. The technique for exploring sea level from analysis of transgressive and regressive stratigraphic sequences goes back to the last century. It was Chamberlin (1909) however who postulated that global transgressions and regressions would provide a natural framework for stratigraphic subdivision and correlation. Stille (1924, 1944) pursued this same idea independently and constructed transgression–regression curves based on stratigraphic data. Grabau (1936) and Umbgrove (1939, 1947) produced similar sorts of curves. The data from which these early curves were constructed were strictly qualitative, and as soon as more quantitative techniques were introduced, they were replaced. The paleogeographic method had begun with Schuchert (1916), who plotted the areas of North America emergent as a function of time. The global paleogeographic atlases of Strakhow (1948) and Termier and Termier (1952) provided a more adequate quantitative basis for estimating the areas of the continents flooded and emergent through time. Egyed (1956a, 1956b) measured the areas on these maps and produced a plot of area flooded versus time in terms of percent continental area.

Wise (1972, 1974) noted some of the problems inherent in attempting to determine the amount of flooding of the continents from paleogeographic maps: (1) the record is obscured by erosion, burial, and metamorphism; (2) continental and upland deposits are more likely to be removed by subsequent erosion than are marine deposits; (3) the base maps have not been palinspastically corrected; (4) global maps include large areas with poorly known geology; and (5) the time intervals represented by individual maps may be very different. Further, it should be added that there is a tendency for the map makers to place the shoreline so that it indicates the maximum flooding which may have occurred during the time interval represented by the map, regardless of whether this was synchronous in all regions. According to Wise (1974), the average number of years represented by successive maps in the Strakhow atlas is 30 million years for the Cenozoic, 40 million years for the Mesozoic, and 60 million years for the Paleozoic. For the Termier and Termier atlas, the average is 10 million years for the Cenozoic, 14.6 million years for the Mesozoic, and 21.2 million years for the Paleozoic. Wise (1972, 1974) concluded that it would be better to use the maps of Schuchert (1955) for North America as a data base and constructed a curve for the percent flooding of North America. The curves of Schuchert, Wise, Termier and Termier, Egyed, and Stille were presented side by side in Wise (1974). In 1974 Wise also presented a refined version of the curve constructed from Schuchert's maps, with time intervals of 7.5 million years for the Cenozoic, 8.9 million years for the Mesozoic, and 6.2 million years for the Paleozoic. He compared this refined curve with the Strakhow/Egyed curve and the Termier and Termier/Egyed curve by superimposing them on the same diagram. The Schuchert map data were then added and averaged to obtain a period-by-period curve that is more readily comparable with those of Egyed. The Strakhow/Egyed, Termier/Egyed, Wise/Schuchert period-by-period and refined Wise/Schuchert continental flooding diagrams are presented in Fig. 6.

The problem of analysis of transgressive–regressive sequences on continental margins has become tractable in recent years with the development of seismic stratigraphic interpretation techniques. Vail and others (1977) have produced a new sea level curve based on analysis of onlap–offlap sequences in continental margin areas. Using data from North America, northern South America, Europe, northern Africa, Japan, Southeast Asia, Australia, and New Zealand, they have constructed a global sea level curve for the Phanerozoic. For the Triassic–Valanginian time they record some 11 cycles of transgression and regression, with an average length of 8.6 million years. Among these, the shortest cycle is that of the Valanginian, representing only about 4 million years. The Hauterivian–Maastrichtian cycles are not shown in detail, having not yet been released by Exxon for publication. For the Cenozoic, they recognize 25 transgressions and regressions before the four rapid cycles of the late Quaternary. The cycles have an average length of 2.6 million years, the shortest being less than 1 million years and the longest, the early Oligocene, being 7 million years. It is interesting to note that for the Tertiary epochs, the average length of cycles (in million years) is markedly different: Paleocene, 2.6; Eocene, 3.1; Oligocene, 3.7; Miocene, 2.2; Pliocene, 1; Quaternary, 0.4. The curve of Vail and others is presented in the same format as the continental flooding curves in Fig. 7.

Calibration of the sea level curves in terms of elevation is a difficult matter. Kuenen (1939) attempted to estimate the eustatic significance of sea level changes

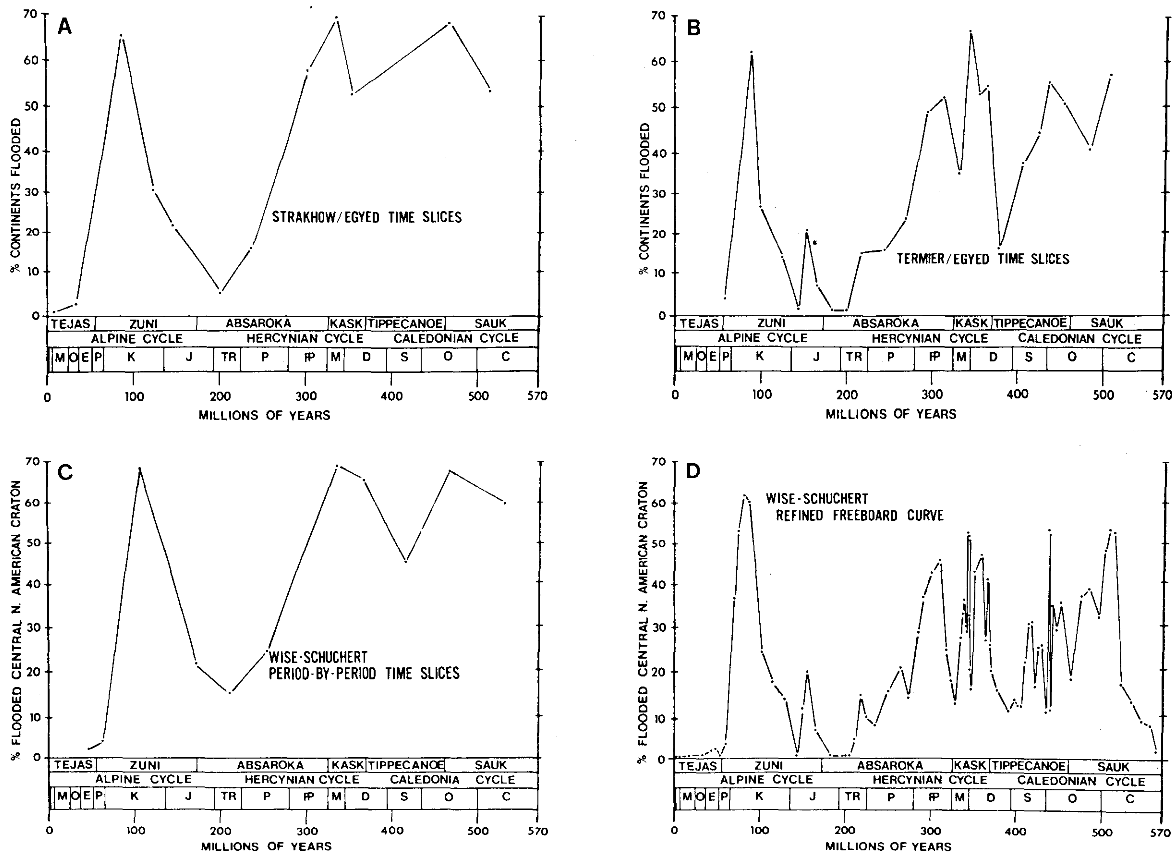

Fig. 6. Percent of continent flooded. (*a*) Egyed's analysis of Strakhow's (1948) paleogeographic maps of the world. (*b*) Egyed's analysis of Termier and Termier's (1952) paleogeographic maps of the world. (*c*) Wise's period-by-period analysis of Schuchert's (1955) paleogeographic maps of North America. (*d*) Wise's analysis of each of Schuchert's (1955) paleogeographic maps of North America.

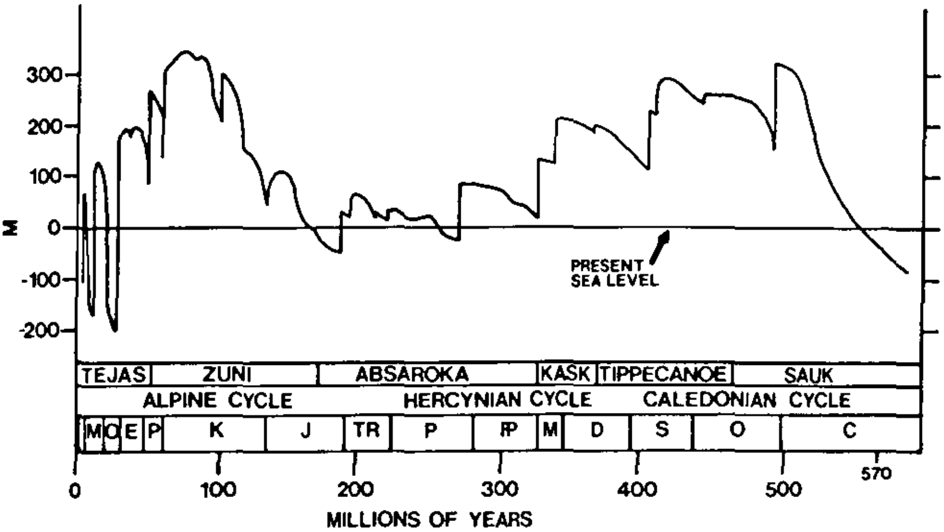

Fig. 7. Vail and others' (1977) sea level curve for the Phanerozoic, calibrated to elevation by Pitman's (1978) late Cretaceous and Cenozoic sea level curve based on change in volume of mid-ocean spreading ridges.

by relating them to the hypsographic curve, noting that at the present time a eustatic rise of 50 m would flood one seventh to one eighth of the land; 100 m would flood one fourth to one fifth, and 200 m would flood nearly one third of the present land area. He assumed that the continents are presently abnormally high because of Tertiary mountain building and estimated that in the past a eustatic rise of 100 m would probably flood one third of the continental area. Hays and Pitman (1973) suggested that the sea level rise of the Late Cretaceous was due to a rapid phase of sea floor spreading and estimated that sea level rose more than 500 m above its present level. Southam and Hay (1977) used the present hypsographic curve to calibrate the Termier and Termier/Egyed curve cited by Wise (1974) with respect to elevation but did not attempt to take isostatic adjustment into account. They arrived at a figure of 330 m for the late Cretaceous transgression, which would be somewhat decreased if isostatic loading of the continent were taken into account.

Sleep (1977), citing evidence from Cretaceous strata onlapping the Precambrian shield in Minnesota, estimated the sea level in the Turonian to be about 300 m above the present.

Utilization of the present-day hypsographic curve to calibrate sea level curves for the past with respect to elevation presents a significant problem. If the Mesozoic–Cenozoic sediments of the continental slopes and rises and of the ocean basins have been offloaded from the continents, at the end of the Paleozoic there must have been some 422×10^6 km^3 of additional material on the continents which, if spread evenly over the present area of the continents, would amount to a layer 2.4 km thick. After isostatic adjustment, this additional material would still require the average elevation of the continents to be at least 370 m above the present value. However removal of the Mesozoic–Cenozoic load from the ocean floor and isostatic adjustment of the continents would lessen the apparent differ-

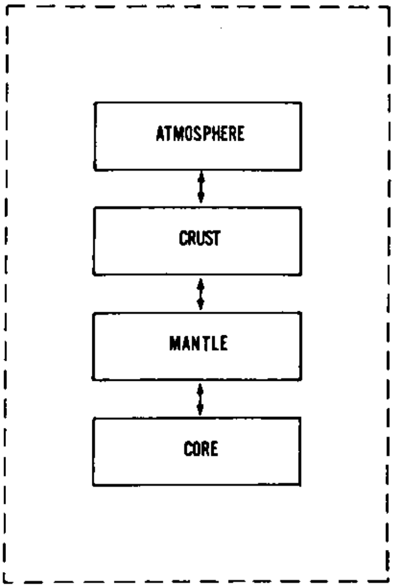

Fig. 8. The Earth as an isolated system.

ence with respect to sea level. In any case, it is evident that the hypsographic curve at the end of the Paleozoic must have been somewhat different from that of today and that calibration of the sea level curve is subject to considerable uncertainty. Fleming and Roberts (1973) have experimented with hypsographic curves for different tectonic situations and give an idea of the range of variation which may be expected with time.

7. Constructing Models for Global and Oceanic Mass Balance

The elements which make up the Earth are distributed in major reservoirs and exchanged among them (see Fig. 7). From this point of view the Earth can be treated as a closed system. The simplest type of model useful in predicting the exchange of material among reservoirs is the box model, which describes the spatially averaged composition of the reservoirs. The validity of the use of such models to predict transients in the system depends on (1) how closely the composition of the individual reservoirs is homogeneous and (2) the time scales of the transient under consideration is long relative to the mixing times reservoirs.

Satisfaction of condition 2 requires different choices of subdivision of the major reservoir depending on the time scale of the transient of interest. For all time scales considered here, the atmosphere will be treated as a well-mixed reservoir. The subdivision of the crustal system into more homogeneous reservoirs within the framework of plate tectonics is shown in Fig. 8. It is assumed furthermore that, during all periods of interest, the mantle and core compositions remain constant in time.

A dynamic description of the Earth system is obtained by balancing the rate at which element i is increasing within reservoir α by the rate at which it is being transported into or out of that reservoir to neighboring reservoirs. To this end let c_α^i represent the spatially averaged concentration of element i in reservoir α and let v_α represent the volume of the reservoir. The geochemical dynamics of the Earth system are described by a set of equations of the form

$$\frac{d(v_\alpha c^i_\alpha)}{dt} = \sum_\beta J^i_{\alpha,\beta} - R^i_\alpha$$

The $J^i_{\alpha,\beta}$ terms are the contributions to the flux of element i from reservoir β to reservoir α through their common surface. These terms are a consequence of material transport by physical processes. The R^i_α term describes the net increase of element i in reservoir α from radioactive decay and production.

8. Geochemical Implications of Plate Tectonics Considered as a System of Reservoirs and Fluxes

A simple model to explain the general distribution of materials in the outer layers of the Earth, derived from the theory of plate tectonics, was presented by Southam and Hay (1977). It can serve a very useful purpose because most tests of the plate tectonic theory have been designed around geometric, geophysical, or paleontologic data. If the basic concept is true, the major chemical fluxes between parts of the system should balance, and the composition of the major observable reservoirs—the oceanic crust, volcanic arcs, and continental cratons—should remain constant through long periods of geologic time. If however the major fluxes between parts of the system are not in balance, the composition of the major reservoirs must be changing with time in a predictable manner.

Figure 9 illustrates the Earth as a series of reservoirs and fluxes, emphasizing the concepts of plate tectonics. The mantle may exchange matter with both the crust and the core. Although processes involving the latter may be important, so little is known about them that they will not be considered here. Layers 2 and 3 of the oceanic crust, distinguished by seismic characteristics, are formed at mid-ocean ridges by material derived from the mantle; they move laterally and are subducted along with the overlying pelagic sediment forming Layer 1 in a Benioff zone. From the Benioff zone, most material will be returned to the mantle, but some will be incorporated through igneous activity into volcanic arcs. Volcanic-arc material may then be welded onto the margins of continents and, through processes of mountain building and reworking, become incorporated into the continent proper. Both the volcanic arcs and the continent proper have lost material back to the deep sea as the pelagic sediment of oceanic crustal layer 1. Thus the Benioff zone is a mixing box for two very different materials—oceanic crust and pelagic sediment. Southam and Hay (1977) developed the model using volume and mass estimates derived from the global hypsographic curve. It is explored here using a different suite of data following the estimates of Ronov and Yaroshevsky (1977).

Figure 10 shows the volumes of the reservoirs for all except the pelagic sediment, for which we used the new estimate given above. The subcontinental volume and mass of Ronov and Yaroshevsky (1977) are included in the continental craton value and the active margin–volcanic arc area is assumed to be equal to the exposed and underwater Mesozoic–Cenozoic geosynclines of Ronov and Yaroshevsky (1977).

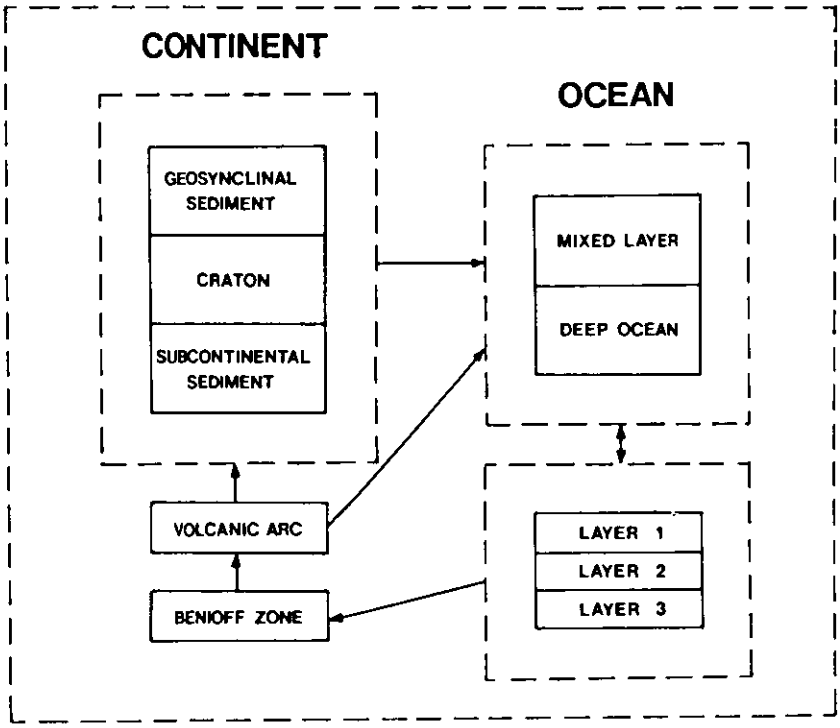

Fig. 9. The Earth's crust as a system.

To estimate the average fluxes between the reservoirs, it is necessary to estimate the age of the ocean floor and the rate of growth of continents. To establish an average age for the ocean floor, the map of the age of the ocean basins by Pitman et al. (1974) was planimetered by Barron and an average age of 104 million years was calculated (Barron, personal communication). This figure has been rounded to 100 million years for convenience. Estimating the rate of growth of continents is more difficult because of the problem of recycling and because the age of the deeper continental crust is speculative. Southam and Hay (1977) considered two limiting values in their calculations: (1) the rate of growth would be zero if the total amount of continental material has not changed with time; (2) the rate would be 49.8×10^{14} g/yr if continental material were initially absent from the Earth and if the rate of growth has been constant through time. The long-term average for the flux of material from continents and volcanic arcs to the pelagic sediment reservoir, layer 1, is estimated by assuming that the age of the pelagic sediments is twice the average age of the ocean floor, namely 200 million years. The flux from each of the land reservoirs cannot be conveniently determined; it is clearly not proportional to area because the relief and hence the sediment supply potential of the volcanic arc is greater than that of the cratonic region.

Figure 9 also shows the reservoirs and fluxes expressed as masses, assuming a constant rate of growth of both volcanic arcs and continental cratonic areas over 4.5 billion years; the net growth rate increment in the volcanic arc is 11.0×10^{14} g/yr; the continental net growth rate is 38.8×10^{14} g/yr.

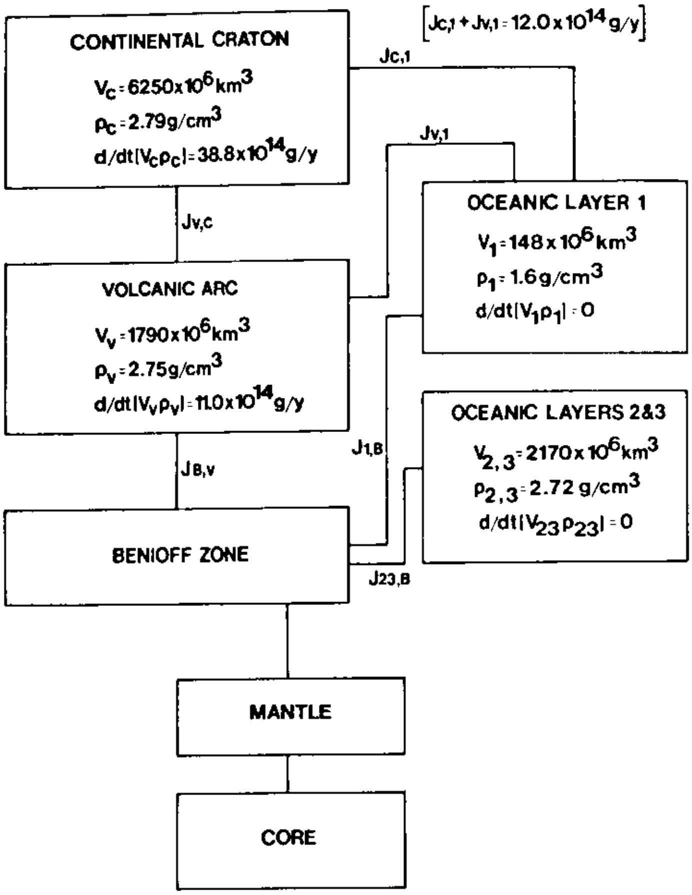

Fig. 10. Volume of Sediment Reservoirs (Except Pelagic Sediments).

9. Determining Chemical Mass Balance for the System

Having estimated the size of reservoirs and the fluxes between them, it is necessary to select values to be assumed for the average composition of each reservoir considered. Because the whole Earth is considered, the more detailed problems of subduction and magma generation as discussed by Gilluly (1971) are avoided. The general mass balance calculation for the crust fortunately does not require an intimate knowledge of the composition of the mantle or of oceanic layers 2 and 3. All of the reservoirs for which a detailed knowledge is needed—volcanic arc, continent, and pelagic—have been intensely sampled with respect to both sediments and the igneous rocks which comprise the bulk of the crust, so that reasonable estimates of their average composition are available. The average compositions used for testing the model, selected for the reasons indicated in the following discussion, are given in Table III.

There are two excellent estimates of the composition of andesitic volcanic arc and active continental margin materials: McBirney's (1969) estimate based on 89 calcic andesites from volcanic arcs and 29 calc-alkaline andesites from active continental margins, chiefly based on samples from Central America; and Markhi-

TABLE III
Composition of Major Crustal Reservoirs in wt % Calculated on CO_2-Free Basis

	Markhinin's (1968) Volcanic Arc	Ronov and Yaroshevsky's (1969) Continent	Poldervaart's (1955) Pelagic Sediment	Southam-Hay's (1977) Poldervaart's Red Clay and Siliceous Ooze
SiO_2	58.1	60.2	35.1	56.7
Al_2O_3	17.1	15.2	9.9	16.9
$FeO + Fe_2O_3$	7.5	6.3	6.3	8.4
MgO	3.4	3.1	2.2	2.9
CaO	7.1	5.5	42.5	8.9
Na_2O	2.8	3.0	1.0	1.5
K_2O	1.2	2.9	1.5	2.7

nin's (1968) estimate based on 427 analyses from volcanic rocks of the Kurils. These two estimates are in remarkably close agreement, never varying by more than 1%. The agreement is even more significant in view of the fact that they were based on mutually exclusive sample sets analyzed in different laboratories. For our calculations we have chosen to use Markhinin's composition because it includes not only the igneous material but also the volcanogenic sedimentary materials and other sediments of the active margin. The above estimates for the composition of volcanic arcs differ somewhat from the estimate for geosynclines given by Ronov and Yaroshevsky (1969, 1977) and by Sibley and Wilband (1977), which indicate a more "continent-like" character. Their estimates are based on data from late Precambrian to Recent rocks, whereas those of McBirney and Markhinin are for areas primarily of Cenozoic age.

For the composition of the continents, the most careful estimates have been prepared by Ronov and Yaroshevsky (1969, 1977) by using a two-layered model for the continental cratonic basement and by including compositional data on the sedimentary rocks overlying the continental craton as determined from calculations of volume and composition of different sediment types.

Southam and Hay (1977) considered a variety of pelagic sediment compositions. For typical modern pelagic sediments, the composition cited is that determined by Poldervaart (1955) from a calculation using the volumetric proportions of the different pelagic chemical sediment types and unpublished analyses by Rubey. Poldervaart's estimate of the proportions of calcareous and noncalcareous components is remarkably similar to that obtained from averaging several thousand carbonate analyses from the Initial Reports of the Deep Sea Drilling Project, which suggested an average carbonate ooze content of about 74% compared with Poldervaart's estimate of 72%. Southam and Hay (1977) presented an estimate of modern pelagic sediment compositions using Poldervaart's volumetric proportions and Sujkowski's (1952) compositions for red clay, radiolarian ooze, diatom ooze, and Globigerina ooze. However we are suspicious of Sujkowski's composition of Globigerina ooze because it indicates 40.2% CaO and only 16.8% CO_2, proportions which seem unreasonable. (If 40% CaO were present, and all the CaO and CO_2 were present as calcium carbonate, the amount of CO_2 should be 31%, or almost twice the amount given by Sujkowski.)

To explore the mass balance before the advent of significant carbonate-secreting plankton, the composition of a pelagic sediment consisting of red clay and siliceous ooze in their present proportions as estimated by Poldervaart (1955) was calculated. Because Poldervaart's siliceous ooze has a composition remarkably similar to his red clay, the composition of the resultant sediment is very close to that of red clay alone. Southam and Hay (1977) also explored the mass balance with a composition calculated as a mixture of Sujkowski's (1952) red clay and radiolarian ooze in Poldervaart's (1955) volumetric proportions. As is discussed in greater detail below, Ronov and Yaroshevsky's (1969, 1977) determination of the average composition of pelagic sediment was based on the areal distribution of the major sediment types and did not take into account the differential sedimentation rates of the different sediment components. Sibley and Wilband (1977) also explored different possible compositions of pelagic sediments. They used as their pelagic sediment 1 that of El Wakeel and Riley (1961) which, like that of Ronov and Yaroshevsky (1969, 1977), was based on areal not volumetric proportions. They then calculated a $CaCO_3$-free pelagic sediment 2 from it, resulting in a composition similar to that of Hay-Southam/Sujkowski's red clay and radiolarian ooze or Hay-Southam/Poldervaart's red clay and siliceous ooze as given in Southam and Hay (1977). Pelagic sediments 3 and 6 of Sibley and Wilband (1977) are red clay and siliceous ooze, respectively, after El Wakeel and Riley (1961). Pelagic sediments 4 and 5 of Sibley and Wilband (1977) are mixtures of the El Wakeel and Riley (1961) average pelagic sediment and the calcium carbonate-free pelagic sediment they derived from it.

Each of these compositions acts as a filter on the flux into the reservoir having that composition, that is, the material flowing into the island arc reservoir has the composition of Markhinin's (1968) andesitic volcanic arc; the material flowing into storage in the continental craton has the composition of Ronov and Yaroshevsky's (1969) continent; and the material flowing from the volcanic arc and continent into layer 1 has the residual composition after material stored in the volcanic arc and continent has been removed. Tables IV and V present the chemical balance in terms of masses of major oxides. In each table the flux from the Benioff zone to the volcanic arc, having the composition of Markhinin's andesite, is shown in the first column. Storage in the volcanic arc is shown in the second column. The amount remaining, which can be passed from the volcanic arc to the continent and to layer 1, the pelagic sediment, is shown in column 3. Storage in the continent is shown in column 4. Predicted loss from the continent and volcanic arc to layer 1 as pelagic sediment is shown in column 5. The amount observed to be stored in the sediment is shown in column 6. Differences between the predicted and observed compositions of pelagic sediment indicate that significant quantities of material are being lost or gained from one or more reservoirs and that the reservoir composition can be expected to change with time.

A special complexity arises in calculating the pelagic sediment mass balance—the presence of CO_2 as a significant component. Although CO_2 does not appear in analyses of island-arc andesites and is only about 1.2% in Ronov and Yaroshevsky's (1969, 1977) estimate of the average composition of continental

TABLE IV

Mass Balance for the System Benioff Zone Volcanic Arc Continental Craton Pelagic Sediment Assuming: (1) Constant Rate of Growth of Volcanic Arc and Continental Craton over 4.5 Billion Years; (2) Pelagic Sediment to Have Always Had Its Present Composition; (3) Compositions and Masses Calculated on a CO_2-Free Basis

	I Benioff Zone to Volcanic Arc	II Stored in Volcanic Arc	III Vocanic Arc to Continental Craton and Pelagic Sediment	IV Stored in Continental Craton	V Predicted Flux to Pelagic Sediment	VI Observed in Pelagic Sediment
	59.1	11.0	48.1	38.8	9.3	9.3
SiO_2	34.34	6.39	27.95	23.36	4.59	3.26
Al_2O_3	10.11	2.84	7.27	6.54	0.73	0.92
$FeO + Fe_2O_3$	4.43	1.43	3.00	2.44	0.56	0.59
MgO	2.01	0.53	1.48	1.20	0.28	0.20
CaO	4.20	1.60	2.60	2.13	0.47	3.95
Na_2O	1.65	0.21	1.44	1.16	0.28	0.09
K_2O	0.71	0.13	0.58	1.12	0.54	0.14

TABLE V

Mass Balance for the System Benioff Zone Volcanic Arc Continental Craton Pelagic Sediment, Assuming: (1) Constant Rate of Growth of Volcanic Arc and Continental Craton over 4.5 Billion Years; (2) Pelagic Sediment to Have Been Noncalcareous; (3) Compositions and Masses Calculated on a CO_2-Free Basis

	I Benioff Zone to Volcanic Arc	II Stored in Volcanic Arc	III Volcanic Arc to Continental Craton and Pelagic Sediment	IV Stored in Continental Craton	V Predicted Flux to Noncalcareous Pelagic Sediment	VI Observed in Noncalcareous Pelagic Sediment
	53.8	11.00	42.8	38.8	4.00	4.00
SiO_2	31.26	6.39	24.87	23.36	1.51	2.27
Al_2O_3	9.20	1.88	7.32	5.90	1.42	0.68
$FeO + Fe_2O_3$	4.04	0.83	3.21	2.44	0.77	0.34
MgO	1.83	0.37	1.46	1.20	0.26	0.12
CaO	3.82	0.78	3.04	2.13	0.91	0.36
Na_2O	1.51	0.31	1.20	1.16	0.04	0.06
K_2O	0.65	0.13	0.52	1.13	0.61	0.11

materials, it accounts for an appreciable part of the mass of pelagic sediment: 22.9% according to Poldervaart (1955). Because CO_2 is atmospherically cycled and is emitted from volcanoes as a gas, it may follow a path different from that of other substances considered in this simple model of the crustal flux system. To avoid this complication, we have reduced the total mass of the pelagic sediment by 22.9%, reducing the average flux over the past 100 million years from 12.0×10^{14} g/yr to 9.3×10^{14} g/yr. Using this value, all of the fluxes have been estimated by calculating compositions on a CO_2-free basis.

Table IV shows the calculated chemical mass balance for the system assuming a constant rate of continental growth over 4.5 billion years. Columns 5 and 6 measure the imbalance in the system and indicate that the system is not in balance with respect to CaO and K_2O for pelagic sediment of modern composition. From these mass balance considerations it is evident that the system "Benioff zone–volcanic arc–continental craton–pelagic sediment" cannot be in equilibrium as long as volcanic arcs are andesitic, as long as the continental cratons have the composition estimated by Ronov and Yaroshevsky (1969, 1977), and as long as pelagic sediment has a composition similar to that which has been accumulating for the past 100 million years. It is clear that CaO is being lost from the continents to pelagic sediments to become ultimately enriched in the oceanic crust or subducted into the mantle. This process, if continued for hundreds of millions of years, must inevitably alter the composition of the continents by depleting them in CaO. During the past 100 million years this process has already significantly influenced the average composition of sedimentary rocks on the continents as well as in the ocean basins. This is not unexpected because Poldervaart (1955) had already commented that pelagic sediment must have had a very different composition prior to the advent of abundant $CaCO_3$-secreting plankton in the middle of the Cretaceous, 100 million years ago.

Hay and Southam (1975) observed that at present the CaO fixed by calcareous plankton and incorporated into pelagic sediment appears to be largely removed by subduction into mantle, conforming to the ideas of Gilluly (1971) and Scholl and Marlow (1974) that most pelagic sediment appears to be subducted. A survey of Spencer's (1974) account of the rocks in Mesozoic and Cenozoic orogenic belts indicates that only a minute fraction of the carbonate lost to the deep sea as *Globigerina* ooze or nannofossil ooze is subsequently obducted into the mountains of active continental margins. With the assumption of continental growth, the system presently operates to remove CaO from the continents and to concentrate it in the oceanic crustal rocks and the mantle. This is a special form of differentiation of the crust, because continental rocks are depleted in CaO relative to oceanic basalts: 5.5% in Ronov and Yaroshevsky's (1969) continent versus 16.7% in their layer 1 basalts. The process would thus effectively increase the chemical contrast. Although pelagic carbonate sedimentation has been operative for only the past 100 million years, it may have already removed about 10% of the total CaO in the continents and has caused 25% of all carbonate sediment to accumulate in the deep sea. As was discussed above, Ronov had noted the marked decrease in dolomite since the Paleozoic and had postulated an overall decrease in carbonate rocks as a whole, from 20% of the total sediment rock volume at the end of the Paleozoic to 5% now. Pelagic sedimentation of carbonates is the most likely explanation for the apparent decrease in carbonate rocks since the Paleozoic indicated by Ronov (1964).

Sibley and Vogel (1976) also noted the calcium dilemma with regard to pelagic sediment and found Hay and Southam's (1975) suggestion of pelagic sediments as a sink for Ca to be a reasonable answer.

Hay and Southam (1975) had assumed that the CO_2 would be driven off as a gas and recycled through volcanoes to drive the weathering system toward leaching of CaO from the continents. However it is possible that about one fourth of the $CaCO_3$ in pelagic sediment might become locked up with the noncalcareous component of pelagic sediment in the form of scapolites having the composition of feldspar + $CaCO_3$ and would be stable under conditions prevailing in the upper mantle. If this is in fact a process which takes place in the Benioff zone, it would serve to gradually remove carbon from the outermost portions of the Earth's crust and to return it to the mantle from which it is thought to have originated by weathering of Scapolites during the Earth's history (Goldsmith, 1976). Calcium carbonate-secreting planktonic organisms would thus have initiated a process which would result in depletion of not only Ca but also C from the Earth's crust.

The mass balance calculations also indicate that only about 0.5% of the K_2O of the continents has been lost to pelagic sediments in the past 100 million years, a depletion rate 20 times slower than that of CaO. Southam and Hay (1977) explored the mass balance for the system under the assumption of no continental cratonic growth. K_2O no longer appeared to be lost to the ocean in significant quantities, but the imbalance for CaO became extreme. Generally the imbalances among major oxides for the no-growth condition were greater than for models which assume a constant rate of growth, so the no-growth assumption was regarded as a less realistic possibility.

Table V shows a mass balance for the system assuming a constant rate of continental growth, but assuming that pelagic sediment consists of red clay which has always been lost to the ocean at its present rate. This is one possible approximation of the pelagic sedimentation system prior to the advent of the abundant calcareous plankton 100 million years ago. Southam and Hay (1977) noted the red clay scheme balances more closely than schemes using modern pelagic sediment. Discrepancies in fluxes based on red clay the composition of Poldervaart (1955) or the Sujkowski (1952) are relatively minor, and the latter appeared to yield a slightly better solution. The implication was drawn that it is possible to grow continents for a very long period of time, billions of years, through the intermediate agency of andesitic volcanic arcs without changing the composition of the volcanic arc, continental craton, or pelagic sediment reservoirs with time. However it is obvious from Table V that the system involving red clay is out of balance with respect to the silica alumina ratio in pelagic sediments.

Southam and Hay (1977) also presented the calculation for the system assuming that pelagic sediment is exclusively red clay accumulating at its present rate and that no continental growth occurred. The solutions were found to be even better than that assuming continental growth at the long-term average rate. However the extremely small residuals are to a considerable extent a function of the fact that all of the flux values are very small for the model based on this assumption. They also explored the mass balance for systems having both continental cratonic growth at the long-term average rate and no continental growth, but with a pelagic sediment consisting of a mixture of red clay and siliceous ooze accumulating at modern rates, hoping to improve the agreement with the observed silica alumina ratio. The solutions were almost as good as those for the system with red clay alone, again indicating that continents could grow by the intermediate agency of volcanic arcs for very long periods of time without changing the composition of the volcanic-arc, continental craton, or pelagic sediment reservoirs. Although this solution is superficially attractive, there does not presently exist any detailed analysis of the data on compositions of either volcanic arc or continental cratonic materials through time. In addition, no *a priori* principle exists requiring the composition of the reservoirs to remain constant. A fruitful line of investigation should be toward determining the average composition of eugeosynclinal volcanic through time, as this may provide special insight into the processes of continental growth and crustal differentiation.

An alternative and more exacting method of exploring the mass balance model is to assume that the fluxes for the most important oxides must be in balance and to predict the rates of storage in the volcanic arc and continental craton. For the conversion of andesitic volcanic arc material to granodioritic continental craton, silica must be concentrated and alumina lost. The implication is that if the mechanism of altering the composition is loss as subducted pelagic sediment, then the pelagic sediment must contain less silica and more alumina than the volcanic arc material. This is in fact true for pelagic red clay of Sujkowski (1952), Poldervaart (1955), and El Wakeel and Riley (1961). It is not true of the opaline components of pelagic sediment or of mixtures of siliceous ooze and red clay in modern proportions. It would be true however if the opaline siliceous component were a mixture consisting only of radiolarian ooze and red clay, both accumulating at their present rates. In fact, 82% of present-day opaline siliceous sediment is diatom ooze and 10% is radiolarian ooze. Since the diatoms are unknown prior to the late Cretaceous, it is reasonable to assume that prior to 100 million years ago only radiolarians contributed to opaline silica sedimentation.

Assuming that continental craton is produced from volcanic arc by loss of red clay to the oceans, it is possible to write equations of the form

$$M_v W_v = M_c W_c + M_p W_c$$

where M_v is the mass stored in the volcanic arc, W_v is the weight percent in the volcanic arc, M_c is the mass stored in the continental craton, W_c is the weight percent in the continental craton, M_p is the mass stored in pelagic sediment, and W_p is the weight percent in pelagic sediment.

Using Table II to obtain the weight percent of the two most abundant oxides, silica and alumina, and assuming the flux of red clay to pelagic sediment to be 1 unit of mass, it is possible to solve for the storage terms for the volcanic arc and continent:

$$\text{For silica: } 0.581M_v = 0.602M_c + 0.554$$
$$\text{For alumina: } 0.171M_v = 0.152M_c + 0.172$$

Simultaneous solution of the two equations yields storage term values for the continental craton of 0.36 mass units per year and for the volcanic arc of 1.33 mass

TABLE VI

Predicted Composition of Pelagic Sediment Assuming: (1) Volcanic Arc Is Transformed to Continental Craton through Loss of Pelagic Sediment; (2) Silica and Alumina Must Balance; (3) Pelagic Sediment Has Composition of Poldervaart's (1955) Red Clay

	I Composition of Volcanic Arc in wt%	II Mass Units Stored in Volcanic Arc (I × 1.33)	III Composition of Continental Craton in wt%	IV Mass Units Stored in Continental Craton (III × 0.36)	V Predicted Composition of Pelagic Sediment (II–IV)	VI Observed Composition of Pelagic Sediment (Poldervaart, 1955 Red Clay)
SiO_2	58.1	77.1	60.2	21.6	55.4	55.4
Al_2O_3	17.1	22.7	15.2	5.5	17.2	17.2
$FeO + Fe_2O_3$	7.5	9.9	6.3	2.3	7.6	9.4
MgO	3.4	4.5	3.1	1.1	3.4	3.1
CaO	7.1	9.4	5.5	2.0	7.4	8.1
Na_2O	2.8	3.7	3.0	1.1	2.6	1.8
K_2O	1.2	1.6	2.9	1.0	0.6	2.8

units per year for each unit of pelagic sediment produced per year. Using these values for all other major oxides, it is possible to predict the composition of the pelagic sediment as shown in Table VI. Using a red clay of Sujkowski's (1955) composition, the storage masses for volcanic arc and continental craton are approximately 1.43 and 0.47, respectively, for pelagic sediment loss of 1 mass unit. As can be seen from Table VII however, the prediction for other oxides is not as good; in particular, Sujkowski gave the combined value of iron oxides as 13.4%, while the model predicts only about half as much, 7.8%. Using the red clay of El Wakeel and Riley (1961), the storage masses of volcanic arc and continental craton are about 1.63 and 0.67, respectively, for pelagic sediment loss of 1 mass unit. Loss of red clay of the composition of El Wakeel and Riley (1961) produces continental craton at twice the rate of loss of red clay of the composition of Poldervaart (1955). Again, however, there is a major discrepancy among the oxides, as shown in Table VIII; CaO and K_2O are particularly poorly predicted. Thus it appears that Poldervaart's (1955) red clay is closest to that predicted by transforming volcanic arc to continental craton and conserving silica and alumina. It is also clear that K_2O is always predicted by any solution to be much less than the amount actually observed.

It is interesting to explore the rate of continental craton and volcanic arc growth suggested by the present-day rate of flux of red clay to the ocean. Poldervaart (1955) estimated that red clay comprises 19.2% of the sediment in the ocean. We have calculated the amount of pelagic sediment as 0.24×10^{24} g, so that red clay would be 0.046×10^{24} g. This represents deposition over 200 million years and the average red clay accumulation during that time would be 2.3×10^{14} g/yr. The predicted fluxes and storage increments are shown in Table IX. Rounding introduces an error of almost 5% in the silica balance, but it is clear that the mass balance is very close for all oxides except K_2O.

TABLE VII

Predicted Composition of Pelagic Sediment Assuming: (1) Volcanic Arc Is Transformed to Continental Craton through Loss of Pelagic Sediment; (2) Silica and Alumina Must Balance; (3) Pelagic Sediment Has the Composition of Sujkowski's (1951) Red Clay[a]

	VII Mass Units Stored in Volcanic Arc (I × 1.435)	VIII Mass Units Stored in Continental Craton (III × 0.470)	IX Predicted Composition of Pelagic Sediment (VII–VIII)	X Observed Composition of Pelagic Sediment (Sujkowski's 1952 Red Clay)
SiO_2	83.4	28.3	55.1	55.1
Al_2O_3	24.5	7.1	17.4	17.4
$FeO + Fe_2O_3$	10.8	3.0	7.8	13.4
MgO	4.9	1.5	3.4	2.1
CaO	10.2	2.6	7.6	6.2
Na_2O	4.0	1.4	2.6	1.4
K_2O	1.7	1.4	0.4	2.1

[a]Compositions of volanic arc and continental craton as in columns I and III of Table IV.

TABLE VIII

Predicted Composition of Pelagic Sediment Assuming: (1) Volcanic Arc Is Transformed to Continental Craton through Loss of Pelagic Sediment; (2) Silica and Alumina Must Balance; (3) Pelagic Sediment Has the Composition of El Wakeel and Riley's (1961) Red Clay

	XI Mass Units Stored in Volcanic Arc (I × 1.626)	XII Mass Units Stored in Continental Craton (III × .666)	XIII Predicted Composition of Pelagic Sediment (XI–XII)	XIV Observed Composition of Pelagic Sediment (El Wakeel and Riley's 1961 Red Clay)
SiO_2	94.2	40.1	54.1	54.1
Al_2O_3	27.7	10.1	17.6	17.6
$FeO + Fe_2O_3$	12.2	4.2	8.0	6.3
MgO	5.5	2.1	3.4	4.6
CaO	11.5	3.7	7.8	1.6
Na_2O	4.5	2.0	2.5	1.3
K_2O	1.9	1.9	0.0	3.7

TABLE IX

Predicted Fluxes and Storage Increments Generated by the Assumptions: (1) Volcanic Arc Is Transformed to Continental Craton by Loss of Pelagic Sediment; (2) Silica and Alumina Are Conserved; (3) Pelagic Sediment Has the Composition of Poldervaart's (1955) Red Clay; (4) Red Clay Accumulates at a Rate of 2.3×10^{14} g/yr (units are 10^{14} g)

	I Flux from Benioff Zone to Volcanic Arc	II Increment Stored in Volcanic Arc	III Flux from Volcanic Arc to Continental Craton and Pelagic Sediment	IV Increment Stored in Continental Craton	V Predicted Flux to Pelagic Sediment	VI Observed Flux to Pelagic Sediment
	6.19	3.06	3.13	0.83	2.30	2.30
SiO_2	3.60	1.78	1.82	0.50	1.32	1.27
Al_2O_3	1.06	0.52	0.54	0.13	0.41	0.40
$FeO + Fe_2O_3$	0.46	0.23	0.23	0.05	0.18	0.22
MgO	0.21	0.10	0.11	0.03	0.08	0.07
CaO	0.44	0.22	0.22	0.05	0.17	0.19
Na_2O	0.17	0.09	0.08	0.02	0.06	0.04
K_2O	0.07	0.04	0.03	0.02	0.01	0.06

It is noteworthy that for any of the above systems in which volcanic arc is transformed into continental craton by the loss of red clay, storage in the volcanic arc is a multiple of the storage in the continental craton. In other words, it is predicted that if red clay is the material lost, volcanic arc areas must be growing more rapidly than continental cratons. It is also apparent that for this model the rate of growth of continental cratons and volcanic arcs is strictly governed by the rate at which red clay is lost to the ocean. Because this rate has been low during the past 200 million years, there is a serious problem involved in attempting to account for the present volumes and masses of crustal reservoirs with this model. First, the increment of continental material generated annually is only 0.83×10^{14} g, which, even over 4.5 billion years, would generate only 0.37×10^{24} g of the estimated 17.44×10^{24} g of continental craton. For the volcanic arcs, 3.06×10^{14} g would be the annual increment, generating only 1.37×10^{24} g of the estimated 4.93×10^{24} g present today. Using a pelagic sediment similar to that of El Wakeel and Riley's red clay, the continental cratonic increment would be 1.54×10^{14} g, providing 0.69×10^{24} g over 4.5 billion years, and the volcanic increment would be 3.75×10^{14} g, yielding 1.69×10^{24} g over the same period.

However if the estimate of the average rate of red clay accumulation over the past 200 million years (2.3×10^{14} g/yr) is low, it is possible for more crust of the proper composition to be generated. One limit may be set by assuming that the andesitic arcs of today, although appearing to be Mesozoic–Cenozoic, in age represent the recycled andesitic material required by the model generated over some longer period of time. This is not as radical an assumption as it may at first appear, because most of the andesitic material is in the Circum-Pacific belt which has probably been an active margin for at least all of Phanerozoic time. If the material were generated over the entire 4.5 billion-year age of the Earth, the red clay accumulation rate would have to be 8.27×10^{14} g/yr (Poldervaart's red clay) or 6.75×10^{14} g/yr (El Wakeel and Riley's red clay), that is, three to four times the estimated average over the past 200 million years. This is not unreasonable since wide-scale vegetation cover of the land has been achieved only during the past 400 million years. Allowing 2 billion years for accumulation of the andesite mass, the red clay sedimentation rates would have to have been 18.5×10^{14} g/yr. These are still less than an order of magnitude greater than that used above and may not be unreasonable estimates for late Precambrian and Early Paleozoic conditions. With such conditions, up to 2.1×10^{24} g of continent could be generated as a byproduct.

An independent estimate of the amount of carbonate-free pelagic sediment which has been produced and lost through subduction has been provided by Sibley and Wilband (1977) using arguments which balance the compositions of average sedimentary rock and average igneous rock. They estimate that 39% of the sediment produced from weathering of igneous rock has been red clay of the composition of El Wakeel and Riley (1961), or 41% if the material is El Wakeel and Riley's pelagic sediment recalculated to be calcium carbonate free. As noted below, our estimate of the existing mass of sediments is 2485×10^{21} g, of which 40.6×10^{21} g is red clay. If the amount of red clay produced is really 39% of all sediment produced but most of it has been subducted, then the total amount of sediment produced through time is 4007×10^{21} g, of which 1563×10^{21} g has been red clay. Only 40.6×10^{21} g remain in the ocean basins, so 1522×10^{21} g have been subducted. This predicts that operating the resulting average accumulation rates would have been 3.5×10^{14} g/yr, over 4.5 billion years, 7.8×10^{14} g/yr, or over 2 billion years. The amount of andesitic volcanic arc material required to be still extant by our model would be 2.5×10^{24} g, and the amount of continental craton produced would be 1.04×10^{24} g. If red clay accumulation had always proceeded at its present rate, it would require 7.5 billion years to produce and subduct the required 1522×10^{21} g.

Even with the assumption of much greater rates of red clay accumulation in the past, it is clear that the great bulk of continental material, at least 15×10^{24} g, must have been generated in the earlier Precambrian by another mechanism. It is necessary to examine in more detail the development of sedimentary rock types to seek evidence of different regimes of crust formation.

Figure 2, modified from Ronov (1964), shows the different proportions of sedimentary rocks through time. It indicates that one major rock type, jaspilite, formed only during the Precambrian. Jaspilite, the rock of the siliceous iron formations, is thought to have been a significant sedimentary component from about 3 to 1.5 million years ago. Its existence may be an important clue to the nature of continental cratonic growth during the Precambrian. Because the content of iron is higher in basalts and tholeiites than in andesite, a process which would generate continental craton material from the more basic rocks would require a byproduct that is more iron rich.

The concentration of K_2O in oceanic basalts is only about 0.2%, or almost 15 times less than in cratonic crust, and suggested compositions for the mantle contain even less K_2O than oceanic basalts. The K_2O concentration in continental cratonic crust can be most readily explained if it is assumed that the primary crustal differentiation took place in the early Precambrian and that only continental growth since then may have followed the path suggested by plate tectonics.

Another clue to the gross crustal differentiation system may be found in the overall masses formed through time, using Ronov and Yaroshevsky's (1977) estimates and assuming that the continental cratonic crust was produced during the first 3.5 billion years of the Earth's history, that the material in and underlying the Riphean–Paleozoic geosynclines was produced between 1 and 0.2 billion years ago, and that material in and underlying the Mesozoic–Cenozoic geosynclines was produced in the last 200 million years. The mass of the continental cratonic crust is 13.21×10^{24} g, of which 0.35×10^{24} g is sediment; assuming the sediment to be mostly Paleozoic, the average rate of growth over 3.5 billion years is 37.7×10^{14} g/yr. The mass of the Riphean–Paleozoic geosynclinal areas is 4.23×10^{24} g, of which 1.01×10^{24} g is sediment; the rate of production over an interval of 800 million years is 84.6×10^{14} g/yr. The mass of the Mesozoic–Cenozoic geosynclinal areas is 4.93×10^{24} g, of which 1.21×10^{24} g is sediment; the rate of production over the interval of 200 million years is 246.5×10^{14} g/yr. The apparent acceleration in the rate of production of continental and geosynclinal mass is evident, but it is not known how much of this is a function of recycling of older materials. Because the rate of growth of the Mesozoic–Cenozoic geosynclinal areas alone is equivalent to the present annual total of sediment transport to the sea, it is clear that a significant part of the growth rate must be nonsedimentary, that is, impacement of igneous rocks or overprinting of older igneous rocks with younger ages due to partial remelting, granitization, or other processes.

10. Effects of the Breakup of Pangaea

A. *Global Hypsography 220 million years ago*

All present continents and some major islands were part of the supercontinent Pangaea at the end of the Paleozoic. The area of Pangaea to the edge of its continental shelf was about 181×10^6 km^2; this includes the area to the edge of the shelf of the present continents (175×10^6 km^2) plus the original area of margin (5×10^6 km^2) attenuated during rifting and subsequently transformed into continental slope as discussed below. It is also assumed that the area of Mesozoic–Cenozoic orogenic belts has remained essentially constant or that their growth is balanced by palinspastic reductions within the older parts of the continent. Assuming that all of the sediment presently in passive margin continental slopes and rises and in pelagic sediments was on the continent at the beginning of the Mesozoic, some 1.76 km of solid, presumably mostly sediment, must be added to the thickness of Pangaea. Adjusted isostatically and assuming a density of 2.7 for the solid phase, the average elevation of the continent would be 320 m above the present average elevation of 744 m. If the sediment had included 22% of air-filled pore space, equaling 500 m of thickness, the average elevation would have been 820 m above the present average elevation, or 1564 m above present sea level. If the pore space were water filled, the average elevation would be 1414 m above present sea level. This does not mean that the average elevation would have been this far above sea level at the time but that it was with respect to present sea level. If isostatic adjustment is taken into account, sea level would have been about 265 m deeper, assuming that the volume of water on the surface of the earth has not changed, that the sedimentary pore space on Pangaea was water filled, and that the ice presently on Antarctica and Greenland was present as seawater. Changes in the

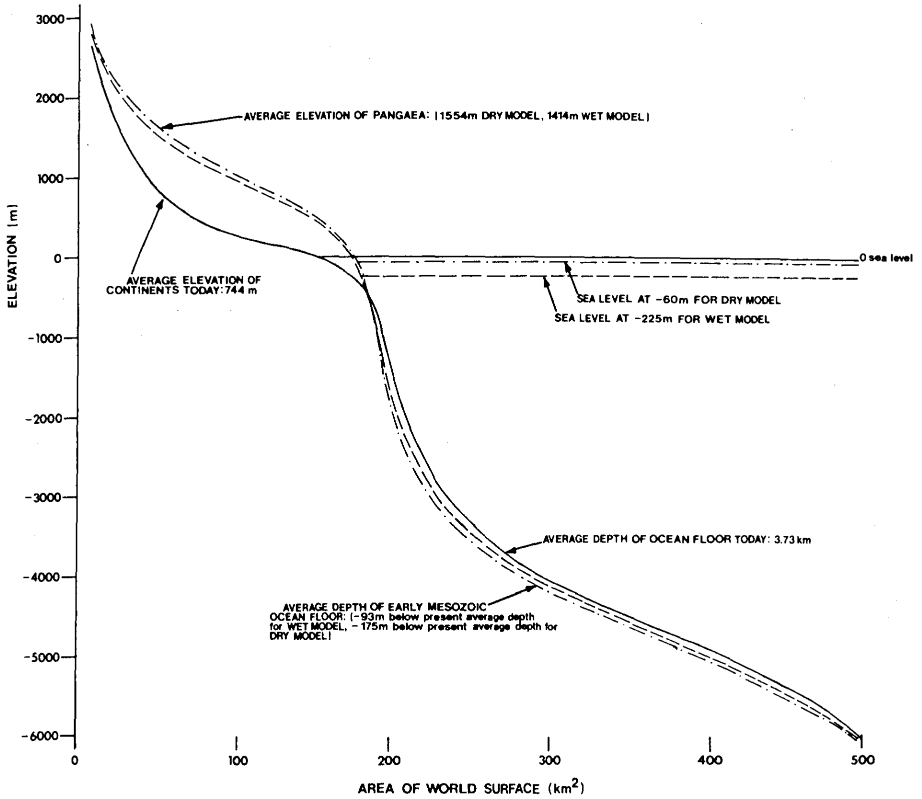

Fig. 11. Present hypsographic curve for the Earth compared with hypothetical hypsographic curves for 220 million years ago. Present hypsographic curve includes Antarctic ice cap but the average elevation given is that for all continents except Antarctica. Melting of Antarctic ice cap would raise sea level to +65 m. Dry model for 220 million years ago assumes no sedimentary pore space filled by water; wet mode assumes all sedimentary pore space filled by water.

size of ocean basins due to formation of continental slopes by tectonic attenuation would raise this level by 40 m to −225 m, assuming no volume changes in the ocean basin due to any other cause such as change in length of spreading ridge.

If none of the additional pore space on Pangaea were water filled, the sea level would have been about 60 m below the present. These relationships are illustrated in Fig. 11, which compares two possible hypsographic curves for the early Mesozoic with that of today. Because of the differences in mass of the materials being shifted, the differences between the early Mesozoic curves and the modern hypsographic curve for land and sea do not match exactly. The curves were constructed by keeping the position of the average elevation of land and ocean over the same point on the areal scale.

The most precise investigation of sea level at the beginning of the Mesozoic is Wise's (1974) analysis of Schuchert's paleogeographic maps of North America, suggesting that during the Triassic the sea level was at least as low as it is at present and that North America was essentially fully emergent. It does not constrain sea level except to a position lower than that of today, and so it does not confirm or reject that hypothetical reconstruction presented here.

As noted by Hay and Southam (1977), the average elevation of a continent is only moderately influenced by mountain ranges, which rarely extend over more than 10% of the continental area. The average elevation is determined largely by the continental craton and platform which constitute the bulk of the continental area. The image of Pangaea is one of a giant continent with broad elevated uplands and an abrupt mountainous margin. The climate must have been relatively dry because of the difficulty of bringing water to the interior of such a large, high continent and because much of the continent lay within what are presently arid zones. Hay and Southam (1977) suggested that very large continents probably have a high ratio of interior to exterior drainage. Extrapolating from information on the present continental ratios of interior to exterior drainage, they suggested that the area of Pangaea having exterior drainage might be as low as 43×10^6 km^2.

The rifting of Pangaea paralleled Paleozoic sutures resulting from continental collision only in the Gulf of Mexico and North Atlantic regions. Other rifts were directed through ancient cratonic areas.

The sedimentary history of the Earth during the Mesozoic and Cenozoic, from 220 million years ago to present, has been dominated by the effects of breakup of this supercontinent. The Mesozoic–Cenozoic passive continental margin shelves, slopes, and rises formed as a result of rifting of Pangaea are estimated to contain 812×10^{21} g, or more than half of the estimated 1605×10^{21} g of extant Mesozoic–Cenozoic sediment. The earliest Mesozoic rifting is that which presumably opened the Arctic basin and the Gulf of Mexico.

Unfortunately these earliest rifting events of the Mesozoic are very poorly known; the rifting history is much more clearly understood for the period since the separation of North America and Africa about 180 million years ago. Since that time, about 64,000 km of rifted margin have been created. Figure 12 shows the increments of passive continental margins formed by rifting over the past 180 million years, following the pattern of breakup which would result from the reconstruction of Gondwanaland by Barron et al. (1978) and the standard reconstruction of Laurasia.

It is widely thought that the breakup of a continental mass by plate tectonic processes follows a distinct pattern involving arching, development of rift grabens, tectonic thinning of continental crust, separation of the continental crust and formation of ocean floor, and subsidence of the new continental margins as they lose heat and move away from the spreading center. To quote Kinsman (1975), "the uplift rupture, and later-subsidence of divergently rifted continental margins are controlled in the first degree by the thermal history of the underlying lithosphere and asthenosphere during these events."

B. Initial Uplift

The manner in which rifting is initiated is still a matter of conjecture. Morgan (1972) has suggested that a more or less linear array of mantle plumes may define the position of an incipient rift, later coalescing to form a spreading center. Burke and Dewey (1973) and Dewey and Burke (1974) have developed the idea of hot spots leading to continental breakup. Baker and others (1972) have suggested that the East African and Arabian–Ethiopian domes are situated over mantle planes and have developed over a 15 to 20 million-year period. Figure 13 shows the

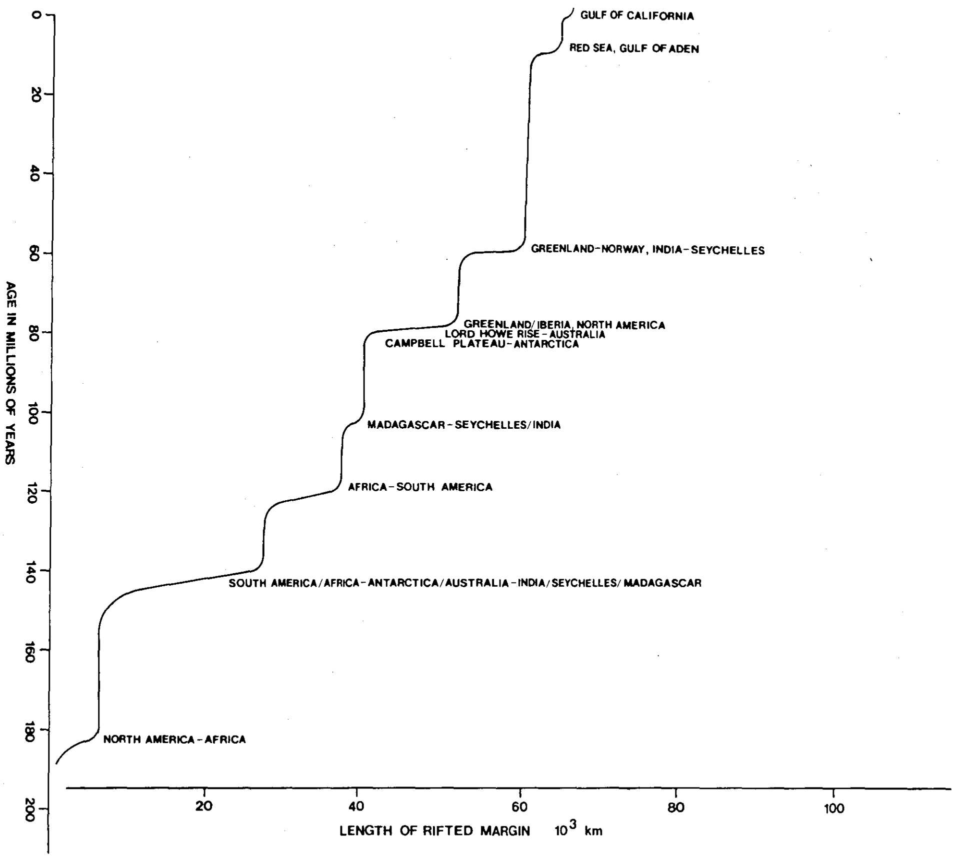

Fig. 12. Increase in length of trailing passive margins through time, 180 million years BP to present.

topography of these domes and compares the real topography with that predicted from modeling. Burke and Whiteman (1973) have described the uplift and rifting processes causing the breakup of Africa. Full development of a swelling of the type of the Atlantic midocean ridge beneath continental crust over a period of 20 million years without erosion could result in formation of an arch about 1.75 km high at the center and 720 km across; most of the elevated bulk of the arch would be within 200 km on either side of the axis. Obviously development of such an arch without erosion is clearly unlikely so that it would not be expected to attain its full potential elevation. It is important to note that the bulk of the sediment eroded from it should be shed to the sides of the arch, and erosion to the original base level could provide as much as 300 km^3 of sediment to each side of the arch for each km of arch length, not taking into account isostatic adjustment to the load loss; with isostatic adjustment this value is increased to at least 900 km^3 of sediment. That complete planation of the arch does not occur is suggested by the shape of the Red Sea, but it is evident that the erosion is a process which results in significant thinning of the continental crust over a wide area and in a major redistribution.

It is also evident that if planation to base level had occurred along all of the 64,000 km of rifted margin, some 58 $\times$ 10^6 km^3 of sediment would have been

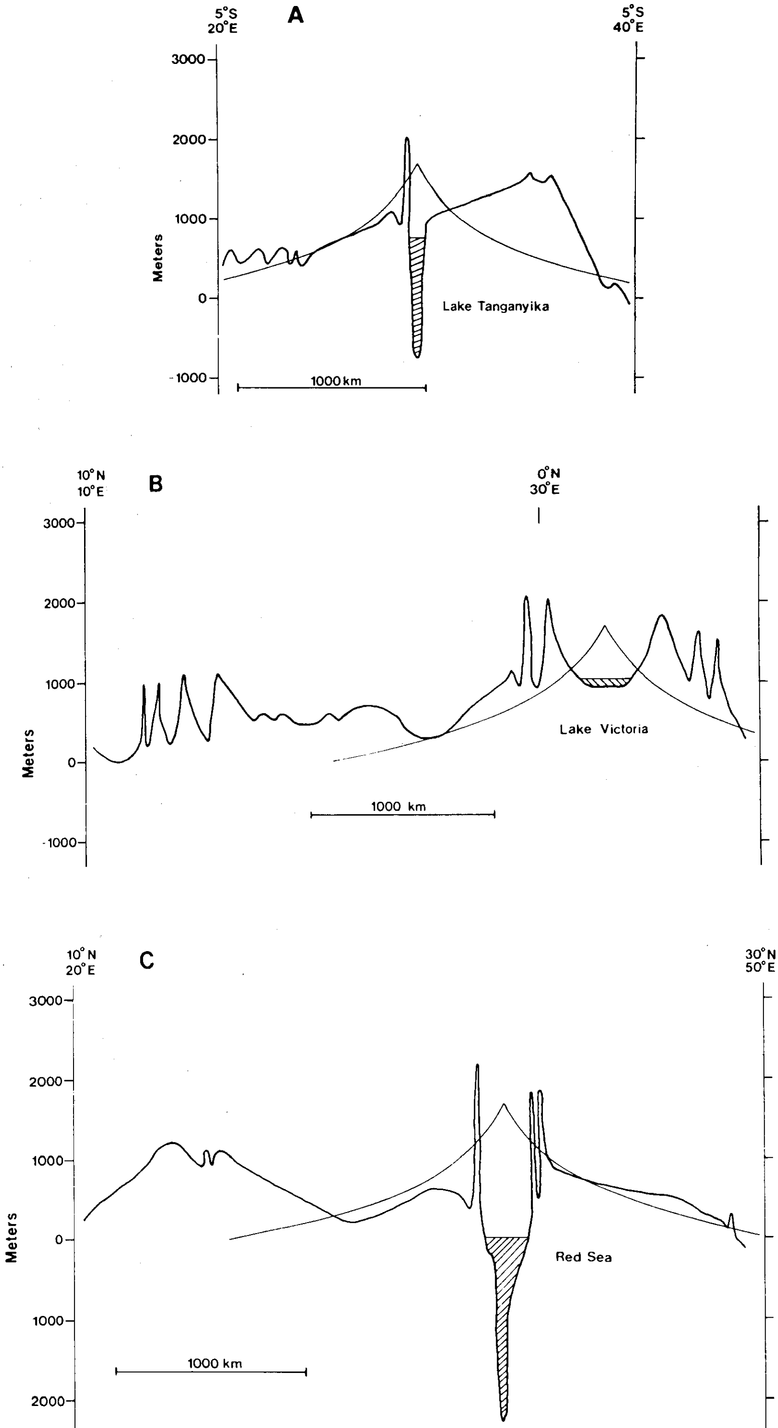

Fig. 13. Sections across the East African Rift and Red Sea Rift. (*a*) Zaire-Tanzania, showing the stratified Lake Tanganyika; (*b*) Cameroon-Kenya, showing Lake Victoria; (*c*) Niger-Saudi Arabia; massif on the left is Darfur Plateau. Profile of a spreading ridge with half rate 1.66 cm/yr and central axis at 1.7 km elevation is superimposed on each profile.

produced. This is essentially equal to the 59×10^6 km^3 of sediment estimated to have accumulated during the Mesozoic and Cenozoic on passive margin shelves.

Kinsman (1975) suggested that largely because of the differing densities of air and water and their loading effect over continental and oceanic crust, continental crust 30 km thick would be expected to subside only 1.7 km during the 100 million-year period when oceanic crust subsides 3.2 km due to thermal contraction. If this is true, uplift of the fully developed arch would be 1.7 km for continental crust 30 km thick. The rifts that fragmented Pangaea affected cratonic areas of a continent which was largely rimmed by a mountain belt, and it seems likely that much of the sediment from the arches would have been deposited on their lower flanks and bordering lowlands, with a relatively small amount transported by rivers to more distant oceanic margin depositional sites. As the arched areas later subsided to form the continental shelves, this sediment became available for erosion and sedimentation on the passive continental margin. The arching and erosion process can result in some crustal thinning, but only in the order of a fraction to a few kilometers.

C. *Graben Formation, Crustal Thinning, and the Lake Phase*

Tension across the uplift causes formation of faults which are thought to originate both at the surface and at depth. Analysis of the Red Sea margins (Lowell et al., 1975) and of other regions suggests that these are generally lystric faults, that is, concave upward, resulting in rotation of the blocks as dilation occurs. The fault grabens become sites of deposition; but because of the nature of the broad regional arching, the grabens have restricted sediment source areas. In the case of the Red Sea, Lowell and others (1975) indicate that the lystric faulting occurred along a band at the center of the arch about 240 km wide. Through lystric faulting, the width was increased to 460 km before actual separation of continental material and exposure of oceanic crust. As a result, the crust was thinned to almost half its original thickness, and tectonically thinned continental margin crust 230 km wide was formed on each side of the Red Sea. The width of this tectonically thinned continental margin may be due to double rifting, that is, rifting at two parallel sites. Burke (1976) has discussed development of the graben associated with the rupture leading to formation of the Atlantic. Kinsman (1975) suggested the tectonically attenuate strip along the Atlantic continental margins to be about 60 km wide and the supracrustally thinned slip, thinned by erosion, to be 140 km wide and to correspond to the continental shelf. More recent work on the Atlantic margins suggests a wider band of tectonically thinned margin, and inclusion of the coastal plain suggests a much wider band of erosionally thinned margin.

Assuming that the continental shelf is underlain by erosionally thinned crust and the continental slope be tectonically attenuated crust (Kinsman, 1975), it is possible to use the present area of the continental slope in regions rifted during the past 180 million years to calculate the average width of erosionally and tectonically thinned margin which has been produced. Using information from Menard and Smith (1966), it can be shown that the area of present continental shelf along the 64,000 km of passive continental margin formed during the past 180 million years is 9.07×10^6 km^2, so that the average shelf width is 108 km. The area of the continential slope (area of slope and shelf below 200 m) is 9.68×10^6 km^2, so that the average slope width is 11 km. Obviously the width of the continental shelf in this calculation is too low to correspond to the average width of erosionally thinned crust. It is better approximated by dividing the total area of passive margin coastal plain and shelf given earlier, 31.8×10^6 km^2, and by the total length of Mesozoic–Cenozoic margins, about 100,000 km, which gives an average width of about 320 km. The average width of the slope obtained, 115 km, appears to be reasonable. Following the proportions of original width versus attenuated margin width determined for the Red Sea by Lowell and others (1975), it may be concluded that tectonic crustal thinning since the beginning of the Mesozoic has enlarged the area of the continental blocks (continent and shelf and slope) by about 5×10^6 km^2 and conversely reduced the area of the continents as measured to the edge of the shelf by about 6×10^6 km^2.

The rift graben becomes a site of special sedimentation processes. Initially formed above sea level, it may not be invaded by the ocean proper until several tens of millions of years have passed. Depending on the local climatology, the rift graben may be the site of swamps, freshwater lakes, salt lakes, or broad alluvial or fluvial deposits. Three typical rift grabens which are well known today are the Rhine Graben bordered by the Vosges and Schwarzwald and accumulating fluvial deposits; the East African Rift, with its large freshwater lakes; and the Dead Sea Rift, which has a significant transcurrent faulting component but which contains a salt lake (the Dead Sea), a freshwater lake (the Sea of Galilee), and a swamp (the Hula Basin).

Rifting in tropical and temperate areas is likely to produce large freshwater lakes. The geology of the East African Rift has been described by Baker et al. (1972). Some of the lakes which occupy the rift (Tanganyika, Malawi, and Kivu) and their sediments have been described by Degens et al. (1971, 1973) and Degens and Stoffers (1976). The tropical lakes tend to be thermally stratified and develop a well-defined pycnocline which results in anaerobic manganese and iron. Sediments can be especially rich in organic carbon, up to 10% in some sediments from Lake Tanganyika. The lake is thought to be in excess of 10 million years old, having been formed in the Miocene, so that it has had a very important long-term influence on sedimentation in the graben. The width of the rift graben is typically 30 to 50 km, and the depth of sediment accumulation may be up to 10 km. Figure 14 shows the cumulative accumulation of terrigenous and lacustrine sediments in early rifts, assuming an average depth of 5 km for the sedimentary fill prior to marine invasion. Assuming an average width of 40 km, an average sediment accumulation of 5 km, and an average organic carbon content of 5% during the freshwater lake stage, the content of organic carbon may be 25×10^{15} g per km of length. If this process were common to all developing rifts of the past 180 million years, the lake stage sediments would contain 8.0×10^{20} g of organic carbon. Since the average concentration of organic carbon in sediments is thought to be 0.3% and the total mass of sediment has been estimated at 2485×10^{21} g, the total amount of organic carbon in sediments would be 7.46×10^{21} g, that is, about 11% of the total organic carbon might be concentrated in only 0.3% of the area of the Earth's surface. Undoubtedly these figures are too high to be realistic, because

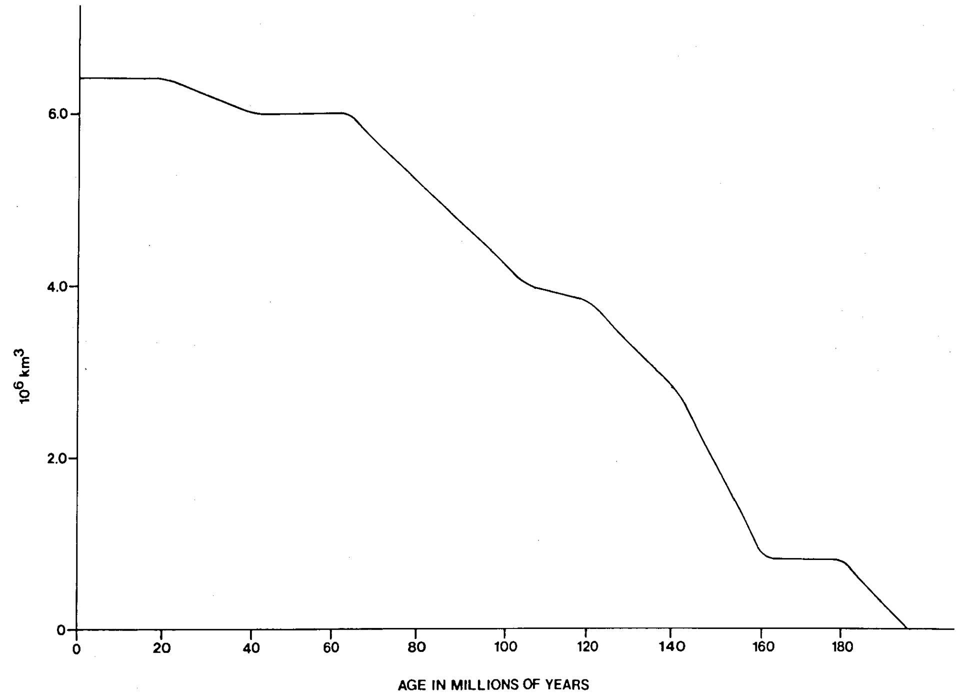

Fig. 14. Increase in volume of terrigenous and lacustrine sediments in the global rift system, 180 million years BP to present.

they assume that anoxic lake deposits would make up the bulk of all early rift graben sediments. Nevertheless they serve to indicate the potential of grabens for accumulation of organic carbon.

The steep-sided walls of the graben generate considerable quantities of coarse sediment which, along with the shore sands and deltaic sediments associated with large lakes, may form porous and permeable reservoirs in close juxtaposition to the rich potential source rocks of the anoxic lake basins, creating unusually favorable conditions for the accumulation of petroleum.

At present the thermally stratified lakes are restricted to the tropics; temperate lakes experience annual overturn and tend to remain relatively well oxygenated. However most of the rifting of Pangaea took place during the Mesozoic when the Earth had an unusually equable climate with tropical conditions much more widespread than at present.

Another example of a young rift is the Dead Sea system. An early invasion of the rift by an arm of the sea from the Mediterranean, through the Valley of Jizreel, resulted in the accumulation of 4 km or more of salt (Zak, 1978). Since isolation from the ocean, the salt has been working its way back to the surface as diapirs (Mt. Sdom, Lisan peninsula), and its exposure on the surface has ensured that a saline lake is present in the graben except during times of unusual dilution by rainfall (Bentor and Vroman, 1957; Neev and Emery, 1967; Begin et al., 1974). The Dead Sea area must once have been the site of unusual accumulations of organic carbon because blocks of asphalt (Amit and Bein, 1978) are sometimes found floating on its surface, washed ashore or coming from unknown sources as well as encountered in drilling. The Sea of Galilee is a freshwater lake, although presently with its surface 200 m below sea level. The Hula basin, at the northern end of the rift system, was occupied by a shallow-water lake, swamps, and bogs, with corresponding large accumulations of organic carbon.

The Dead Sea rift scarps are bordered by fanglomerates and accumulations of coarse clastic material with high porosity and permeability, but salt tectonics have greatly disturbed the older sediments.

D. *Marine Invasion of the Rift*

As the rift broadens and deepens, the central graben depression will sink below sea level and eventually the ocean will flood the graben. In this stage it is common for thick accumulations of evaporites to occur. In the Red Sea, seawater is thought to have entered from the north, through the Gulf of Suez (Kinsman, 1975) and to have resulted in the accumulation of 7 to 8 km of evaporites near the present Red Sea shoreline (Hutchinson and Engels, 1972). Accumulation of salt to a thickness of 3 to 4 km continued as the thinned continental margins separated and ocean crust was emplaced.

The phenomenon of evaporite accumulation as the graben reaches sea level appears to be widespread. Evaporite extraction by the Gulf of Mexico, North Atlantic, South Atlantic, and Red Sea are now relatively well known as indicated in Fig. 15. Rifting to form the Indian Ocean may not have resulted in evaporite extraction because the rifts occurred at higher latitudes. It is important to note that the chance for evaporite extraction is enhanced by the nature of the rifting process. The shortest rifted margins we know of (Red Sea) are in the order of 1000 km, and transform fault offsets tend to result in a series of semiisolated basins in the early opening phase. The narrow basins are paralleled by the mountainous walls of the arched continent which may be 2 to 3 km above sea level. Streams feeding the basins are restricted in their headwater areas to the region immediately surrounding the basins, for the broad regional uplift directs most of the runoff away from the basin and causes larger rivers, such as the Nile, to flow parallel to the basin. Finally, air entering the depression, unless entering along the length, must first pass over the bordering mountains where its moisture will be extracted. As the air descends, it is adiabatically heated and its capacity to accept water vapor is greatly enhanced.

It appears therefore that early rifting tends to create conditions first for the accumulation of organic carbon and potential petroleum reservoir rocks and then for evaporite deposition to form an excellent seal for a potential petroleum reservoir. The potential for the rifting process to provide a sink for organic carbon and then a sink for $CaSO_4$, NaCl, and KCl is a very important factor in understanding the chemical balance of the Earth's surface over the past 200 million years. A single event, the extraction of 4×10^6 km^3 (100×10^{20} g) of evaporites between 120 and 100 million years ago, would have altered the salinity of seawater by about 20%. The coincidence of the end of this extraction with the abundant appearance of calcareous plankton as fixers of carbonate and contributors to pelagic sediments is intriguing. Was this lowering of salinity which resulted in the carbonate secreting planktonic Foraminifera and coccolithophores populating the

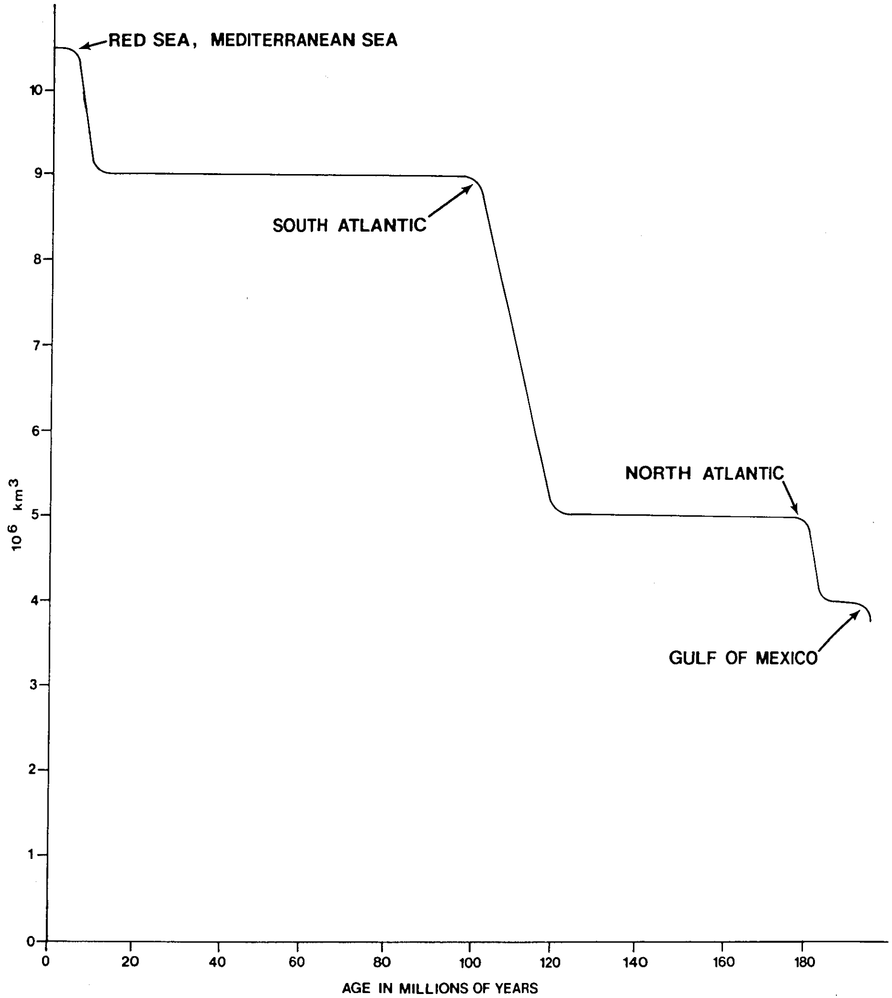

Fig. 15. Extraction of evaporites from the ocean, 200 million years BP to present.

surface waters? Figure 16 shows a first attempt to reconstruct part of the salinity history of the ocean. Chloride ion was selected because it is removed in large quantity in evaporite basins and because it has a large residence time in the ocean. The chloride ion in river water is assumed to come from atmospherically recycled salt, from volcanic sources, and from dissolution of salt deposits exposed to circulating ground water. The lower curve assumes that the supply of chloride to the ocean has remained constant, while the upper curve assumes that it varies proportionally with the supply of $CaCO_3$ to the ocean, as documented from study of the Deep Sea Drilling Project data.

E. Separation

The first appearance of oceanic crust marks the moment of continental separation and is the first event which can be precisely dated by magnetic anomaly studies. It does not necessarily mark a change in sedimentation regime. In the Red

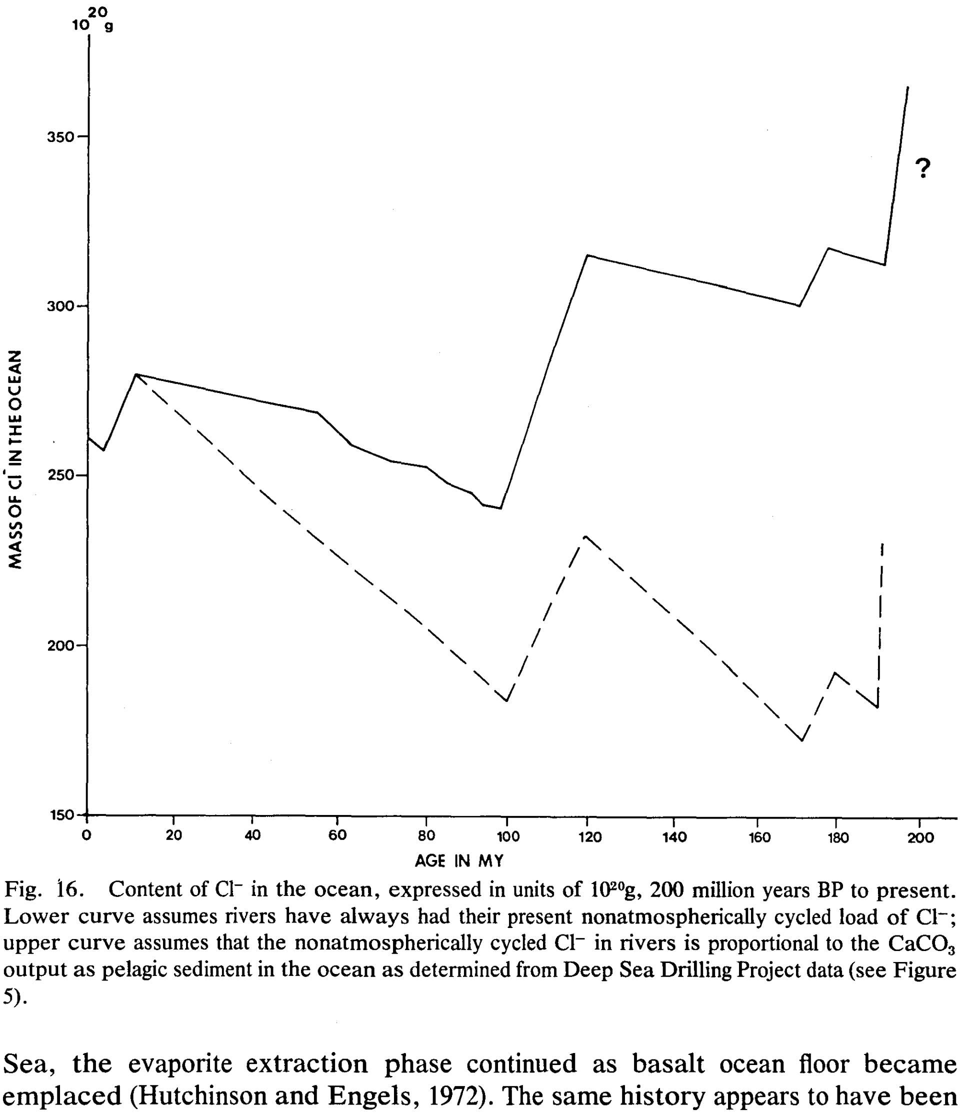

Fig. 16. Content of Cl^- in the ocean, expressed in units of 10^{20}g, 200 million years BP to present. Lower curve assumes rivers have always had their present nonatmospherically cycled load of Cl^-; upper curve assumes that the nonatmospherically cycled Cl^- in rivers is proportional to the $CaCO_3$ output as pelagic sediment in the ocean as determined from Deep Sea Drilling Project data (see Figure 5).

Sea, the evaporite extraction phase continued as basalt ocean floor became emplaced (Hutchinson and Engels, 1972). The same history appears to have been characteristic of the South Atlantic, which was isolated for 20 million years, after initial separation from connection with the open ocean, by narrow straits along transform faults.

F. *Subsidence*

As separation proceeds, subsidence of both the tectonically and erosionally thinned continental margin occurs. The relation of this process to thermal subsidence of the oceanic crust has been postulated by Sleep (1971) and Kinsman (1975). Sleep (1971) followed the notion of Dietz (1963; 1965) and Rona (1970) in believing that a strong coupling must exist between the continental margin and the ocean basin. Kinsman (1975) argued that continental margin subsidence and that of the adjacent oceanic crust, although both responding to thermal subsidence, are not coupled. He presented theoretically derived subsidence curves illustrating the process. Kinsman suggested that the "continental terrace," the erosionally thinned portion of the continental margin, subsides to sea level about 30 to 50 million years after separation. There are now a number of investigators who have explored a variety of models for sedimentation on passive trailing margins (Sheridan, 1969; Bott, 1971; Walcott, 1972; Rona, 1973; Falvey, 1974; Watts and Ryan, 1976; Whitten, 1976). These need to be explored not only in the context of subsidence due to thermal compaction but also taking into account changes in sea level, isostatic adjustment, and sediment supply.

As the continental terrace subsides to sea level, the regional slope reverses from tilt away from the rifted margin to tilt toward the rifted margin. Drainage patterns reverse, and the sedimentary wedge originally derived from the prerift arch can become an important source of sediment to the subsiding margin. Isostatic adjustment to loading of the shelf by accumulating sediment causes it to subside more rapidly than thermal decay alone would require. Increase of subsidence rate due to isostatic load depends upon the density of the sediment and its compaction history as loading proceeds. If the average density of the sediment is 2.2, which assumes a porosity of 30% and a solid phase of density 2.7, isostatically adjusted subsidence is three times that without sediment load. For example, if the shelf break is maintained at 100 m below sea level and sea level remains constant, thermal subsidence of 500 m must be accompanied by 1500 m of sediment deposition in order to maintain the shelf profile and retain the shelf break at 100 m below sea level. It is obvious that both compaction and isostatic subsidence are important in modifying the effects of thermal subsidence; practical techniques for correction for these factors have been presented by Van Hinte (1978).

Subsidence of oceanic crust away from the midocean ridge crest with declining heat flow has been described by Sclater and Francheteau (1970) and McKenzie and Sclater (1971). It has been noted by Sclater et al. (1971) that a simple relationship exists between depth and age for ocean floor younger than 80 million years. Parsons and Sclater (1977) have found that for ocean crust older than 1 million years and younger than 80 million years the relative subsidence of the oceanic crust, h_{oc}, follows within ±300 m the empirical relation

$$h_{oc} \approx 350 \sqrt{t} - 200$$

where t is expressed in millions of years (see also Sclater et al., 1977).

Kinsman (1975) observed that water loading makes a significant contribution to the oceanic crust subsidence curve. The additional load of 3.2 km of water added to oceanic crust over 90 million years depresses the curve about 0.97 km below its level if subsidence involved continental crust exposed to air. He postulated that subsidence of continental crust beneath an air load would only be 1.7 km over a 90 million-year period if there were no sediment loading. His argument is based on the assumption of a mantle density of 3.3 under a spreading ridge or continental arch and that pressure equilibrium is attained at 100 km depth. He also observed that subsidence first under air and then water load would result in a sudden increase in subsidence rate as the surface passes beneath the air–water interface.

From the existence of continental rises and a shelf break, it is evident that the supply of sediment to most passive margin shelves has exceeded the accumulation capacity of the shelf as it undergoes thermal subsidence and isostatic adjustment to loading. Assuming that the North American shelf with an average maximum sediment accumulation of 6 km is typical, a general curve for the subsidence history of passive margin can be plotted. The age of the sedimentary column varies according to the time at which rifting was initiated. For the major passive continental margin segments formed in the past 180 million years the predicted age distribution for sediments in the average maximum accumulations is given as hypothetical stratigraphic columns in Fig. 17. For wells of North America described by Van Hinte (1978), the agreement of age–depth relationships is close. The total volume of sediment accumulation on continental shelves from 180 million years to the present is indicated in Fig. 18.

Seismic profiles off North America suggest the presence of extensive barrier reefs during the lower Cretaceous, with a shelf margin located more seaward than the present one. At the time the reefs formed, the Grand Banks were at about 35°N, slightly north of the present latitude of Bermuda. Migration of reef growth to the more southerly portion of the North American margin with time is in part a function of the rotation of the margin which has shifted the Grand Bank northward to 45° and in part a function of the more sharply defined climatic gradients which exist today.

The materials that accumulate on the shelf are chiefly terrigenous detritus and shallow-water carbonates. The detrital materials are introduced at river mouths and point sources and are either spread over the shelf by currents or, if the supply exceeds the dispersive capacity of the shelf currents, spill over the shelf break to accumulate on the slope, rise, and abyssal plains. The detrital supply is governed by the vagaries of paleogeography, although it is interesting to note that the major rivers of the world debouch over passive margin shelves. Most of the length of passive margins is characterized by relatively small rivers carrying relatively low detrital loads. Nevertheless the present-day detrital load of rivers so vastly exceeds the space being created on the shelves by subsidence that it is obvious that about 95% of the detrital load must be eventually bypassed to the deep sea. At the moment, the late Pleistocene–Holocene rise in sea level has flooded the shelves, creating an anomalous condition which is discussed in the section on the effects of sea level changes.

Carbonate accumulation occurs by drawing upon the reservoir of dissolved carbonate in the ocean. Although ultimately supplied by rivers, the carbonate is not introduced at point sources but is available along the entire margin system. Whether carbonate accumulation or detrital accumulation will occur at a given site depends on whether or not conditions are appropriate for biologic fixation of carbonate as reefs, banks, or platform deposits. Direct precipitation from seawater may occur under unusual circumstances, but the overwhelming bulk of carbonate is biogenic, and at the present time carbonate accumulations in excess of 30% of the total sediment are limited to the tropical and subtropical regions between 35°N and 35°S. Assuming that these latitudes have been the limits of carbonate accumulation for the entire Mesozoic and Cenozoic, it is possible to estimate the availability of shelf space for carbonate accumulation through time. As noted in the section on estimates of sedimentary volumes, masses, and composition, the carbonate accumulations that appear in seismic sections of the continental margins plus the carbonate accumulation in deep-sea sediments suffice to make up the deficit suggested by Ronov (1964).

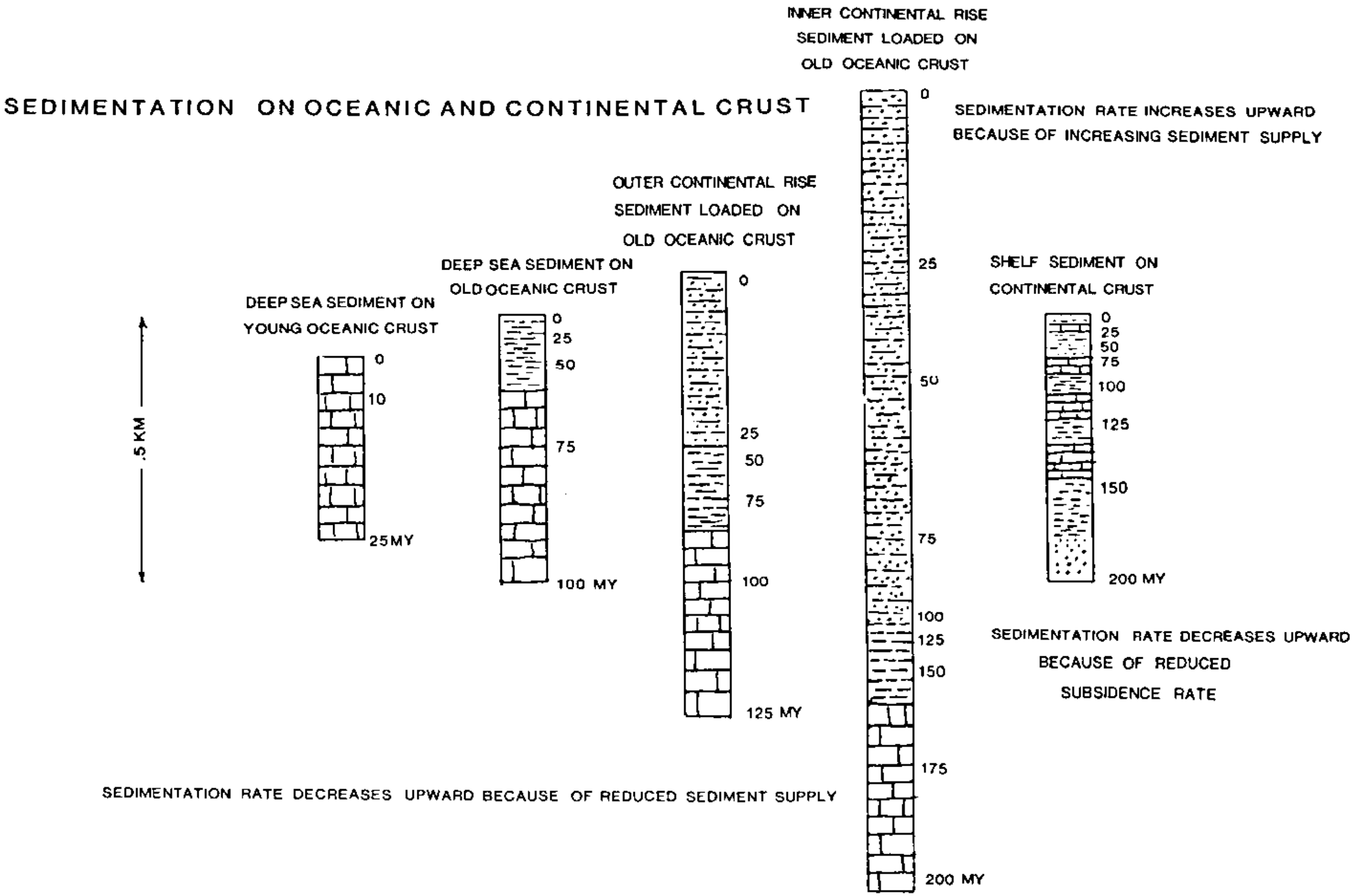

Fig. 17. Idealized stratigraphic columns showing sedimentation rate and type for the mid-ocean ridge, continental rise, and continental shelf. The rate of pelagic sediment accumulation in the ocean decreases as the crust subsides into the zone of $CaCO_3$ compensation; the rate increases as rise sediments prograde basinward; the rate of shelf sedimentation decreases as the subsidence rate declines with the rate of thermal contraction.

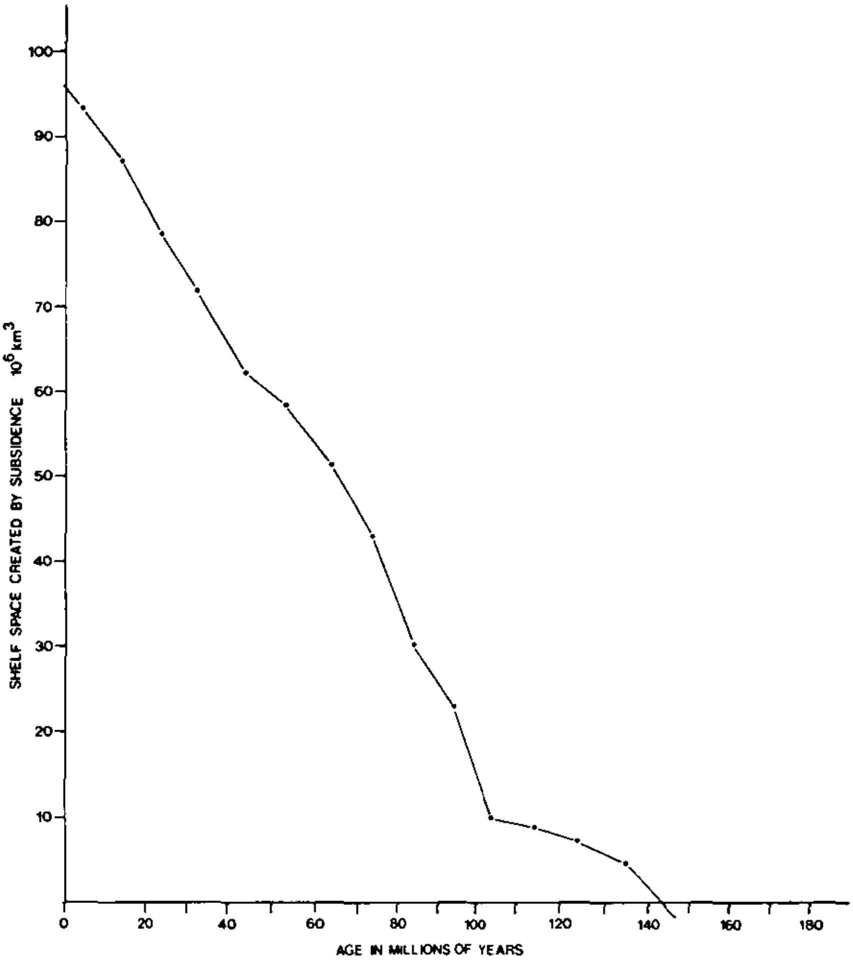

Fig. 18. Shelf space available for sedimentation, created through thermal contraction of passive trailing margins indicated in Figure 20.

11. Eustatic Changes of Sea Level

Sea level changes can be due to a large number of causes, and the implications of plate tectonics are highly significant for the interpretation of the sea level record. The total volume of water on, and associated with, the surface of the Earth has been estimated by Garrels and Mackenzie as 1.72×10^{24} g, distributed among reservoirs as follows: 1.37×10^{24} g in the oceans, 0.33×10^{24} g in pore spaces in sediments, 0.02×10^{24} g as ice, 0.03×10^{21} g as fresh water in lakes and streams, and 0.013×10^{21} g in the atmosphere. The position of sea level is in part a function of the partitioning of H_2O between these reservoirs. Next to the oceanic reservoir, the largest reservoir is in the pore spaces of sediments. Our new estimates of volume and mass of sediments indicate that the volume of sedimentary pore space available to water is 176×10^6 km^3, which could accommodate 0.18×10^{24} g of water. Space for water also exists in the midocean ridge system. From thermal considerations, the flux of water through the ridge system has been estimated at

1.5×10^{17} g/yr (Holland, 1978), but the actual space is not well known. If the area of the ridge through which water circulates is estimated as 120×10^{6} km^2 with an average thickness of 4 km and 10% pore space, the volume of water would be 48×10^{6} km^3, corresponding to 0.05×10^{24} g. This volume will vary with size of the midocean ridge system.

Unfortunately there is no fixed gauge on the Earth to provide an absolute reference against which sea level can be measured. In the following discussion sea level changes are expressed in terms of meters with reference to a hypothetical fixed gauge without taking isostatic compensation into account. The nearest approximation to a fixed gauge in the ocean is probably a small island which responds as an integral part of the ocean floor. Observed from such an island, a sea level rise of 100 m would appear to be 100 m; with subsequent isostatic adjustment the same rise observed from the nonflooded part of the continents would be 70 m. The time required for isostatic adjustment is geologically short, a few tens of thousands of years (Chappell, 1974). As noted in the section on documentation of sea level changes interpreted from the paleogeographic record of continental flooding and seismic studies of continental margins, sea level records are initially expressed as shoreline positions and are subsequently calibrated by means of a hypsographic curve or other data to be expressed in terms of elevation or depression with respect to present sea level. The shoreline position is determined not by sea level alone, but by the rate of sea level change and the rate of delivery of sediment. If the sediment supply were excessive, the shoreline could migrate seaward even if sea level were rising. Conversely it is possible to imagine that if sediment supply from inland were very small, coastal erosion processes could move the shoreline landward even if sea level were falling. To evaluate the observed record of sea level change, it is necessary to explore all of the possible causes and to be aware of the isostatic feedback mechanism.

A. Possible Very Long-Term Trends

Addition of "Juvenile Water" from the Interior of the Earth

Rubey (1951) in his Presidential Address to the Geological Society of America discussed the problem of the origin of seawater and the other "excess volatiles," C, S, N, Cl, He, B, Br, A, Fl, and so on, which are concentrated in the atmosphere, hydrosphere, biosphere, and sediments. Arguing that these cannot be derived simply from the weathering of igneous rocks, he calculated fluxes at which these materials would be delivered from subcrustal sources if their rate of supply were constant through geological time. For water, the rate would be 3.7×10^{14} g/yr, which is five orders of magnitude less than the 0.32×10^{20} g/yr delivered by streams. If this were a steady addition, it would cause sea level to rise 1 m every million years, or 1 cm/1000 yr.

Subduction of Pore Water

Taking into account plate tectonics, and assuming the amount of sea floor subducted to be equal to the amount of new sea floor formed each year, 1.48×10^{6} km^2, the amount of pore water subducted in pelagic sediments is 2.6×10^{14} g/yr. If water were driven out of the sediments as they are subducted and heated, it would appear as "juvenile water" from volcanic sources. Actually the amount of apparent juvenile water is considerably larger because a probably larger mass of trench sediments from other sources are subducted along with the pelagic sediments. The flux out via sediment and subduction acts to lower sea level at a rate of 0.7 to 1.4 cm/1000 yr.

It is now evident that subduction could supply the water thought to be derived from deep sources and that the net flux could be into the mantle rather than out of it. It is also clear that these processes would not produce any detectable changes other than possible long-term trends over the time scale in the hundreds of millions of years.

Expansion of the Earth

Egyed (1956a, 1956b) related a long-term decline in sea level to expansion of the Earth. Such long-term decline was postulated from the curves showing the area of continents flooded as a function of time, constructed by Egyed from the paleogeographic maps of Strakhow (1948) and Termier and Termier (1952). The average rate of expansion of the Earth's diameter during the Phanerozoic was taken as 0.5 mm/yr. The area of the Earth's surface 100 million years ago would have been 495×10^6 km^2, which is 15×10^6 km^2 less than at present. If all the expansion took place in the ocean basins and the volume of water remained constant, sea level would have been 178 m higher 100 million years ago and would have decreased at a rate of 0.18 cm/1000 yr. This long-term trend is of the same order of magnitude but in the opposite sense of that postulated by Rubey (1951). Although the possibility of expansion of the Earth is not considered seriously by most geologists and geophysicists today (Le Pichon, 1968; Hallam, 1971), some of the modern concepts of plate tectonics were developed by Egyed (1956a, 1956b) and Carey (1956) in the context of an expanding Earth. There is no evidence from physics to decide whether the universal gravitational force G is a constant or decays with time, as suggested by Dirac.

B. Possible Oscillatory Trends with Time Scale 10^7 to 10^8 Years—Volumes of Midocean Ridges

The present length of the midocean ridge system is 54,694 km. The area of ocean crust which has been generated from the midocean ridge over the past 200 million years is about 296×10^6 km^2, so that the average rate of sea floor separation over the past 200 million years would be 2.71 cm/yr, or a 1.35 cm/yr half-rate for sea floor spreading if the length of the midocean ridge had remained constant through time. A midocean ridge spreading at 1.35 cm/yr displaces about 2070 km^3 water for each kilometer of its length.

More than two thirds of the presently observed ridge, 37,474 km, is in the Arctic, Atlantic, and Indian Ocean basins and has obviously been generated within the last 200 million years. It appears that as a general rule midocean ridges bound by subduction zones spread more rapidly than those bound by passive continental margins. The average half rate for spreading centers bound by continental margins, those of the Arctic, Atlantic, and Indian Oceans, is about 1.66 cm/yr (Le Pichon, 1968). The slow rates for the Atlantic seem to have persisted throughout the development of that ocean. Assuming that the length of ridge in the Pacific has remained constant and assuming a half-rate of 1.66 cm/yr, each kilometer of Arctic–Atlantic–Indian ridge displaces 2546 km^3 and the total displacement of the Arctic, Atlantic, and Indian Oceans is 95.41×10^6 km^3, which would have caused a rise of sea level of 265 m in the last 180 million years, or an average of 1.5 cm/1000 yr.

It is obvious that the length of the ridge system is important in changing sea level, but for a detailed account the history of ridges in the Pacific basin must be known with greater precision than is known today. It seems evident that the length of ridge in the Pacific basin 200 million years ago could not have been as great as that of today unless the ridge system contained several triple junctions.

The amount of water displaced by a midocean ridge undergoing thermal subsidence is directly proportional to the rate of spreading. The average of spreading half-rates for the Atlantic and Indian Oceans determined from data of Le Pichon (1968) are the same, 1.66 cm/yr. The average half-rate calculated from Le Pichon's data for the Pacific ridge system, which is bound by subduction zones everywhere except along Antarctica, is 3.95 cm/yr, more than double that of the spreading ridges bound by passive continental margins. Doubling the rate of spreading along the 17,220 km Pacific segment would ultimately displace 104.3×10^6 km^3 of water if the rate remained doubled for 90 million years and cause a rise in sea level of 289 m. If the rate were doubled for 10 million years and then returned to its original value, the volume displaced would produce a sea level rise of 103 m, which would develop over the 10 million-year period but would require 90 million years to decay away completely. If the rate doubled for 20 million years, the sea level rise would be 170 m.

Pitman (1978) has modeled the shape of the sea level curve by a change in spreading rate to determine its shape in greater detail, as shown in Fig. 19.

Pitman (1978) has also presented detailed calculations of ridge volume at 10 million-year intervals from 85 million years ago to the present, taking into account both ridge length and spreading rates along each ridge segment. He noted that calculation of the ridge volume at 85 million years ago requires that the spreading history of each ridge segment be known back to 155 million years ago and that much of the older ocean floor has been subducted. The spreading data for the older ridge segments were estimated by extrapolation and could be significant sources of error.

Pitman's curve for sea level change is given in Fig. 20 together with Vail et al.'s (1977) curve for the sea level changes observed on continental margins. Pitman's sea level elevation was determined by multiplying the raw calculation by 0.7 to take into account isostatic adjustment of sea level with respect to the continent. Vail et al. (1977) used Pitman's curve to calibrate their sea level curve to absolute elevation with respect to modern sea level; this calibration is with respect to an oceanic island behaving as an integral part of the sea floor. The calibration should still be regarded as tentative with regard to continental margins because isostatic compensation of the continental blocks and their change in hypsography with time were not taken in to account.

C. *Emplacement of Volcanic Chains or Aseismic Ridges*

Linear volcanic chains or aseismic ridges, thought to be produced by hot spots in the mantle as the plate moves past, also displace water. Pitman (1978) termed these "thermal welts" and calculated that if the feature had the shape of a

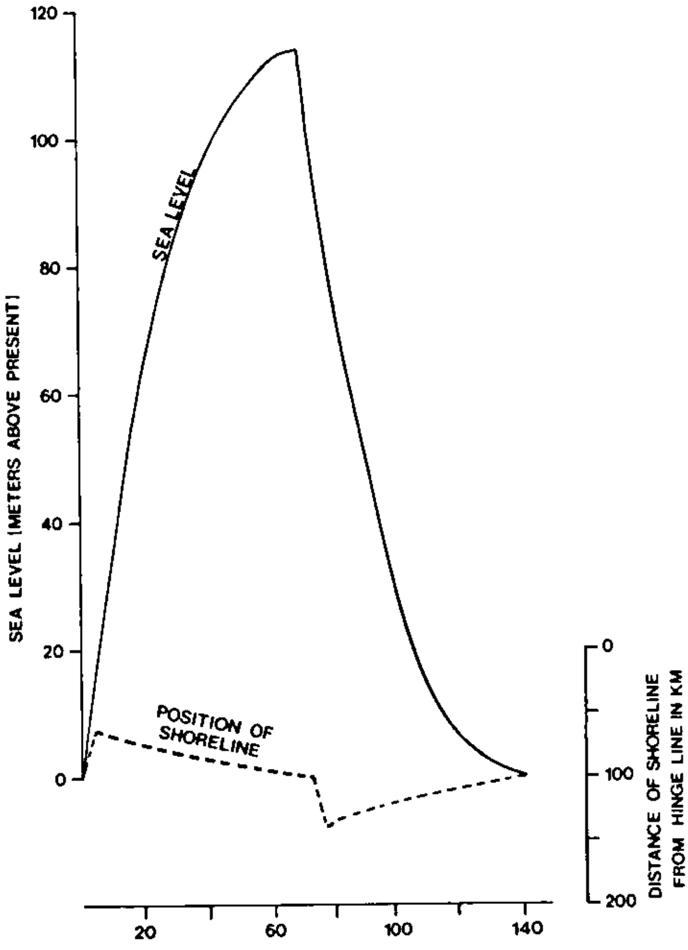

Fig. 19. Pitman's 1978 model of sea level during episode of rapid increase; it is assumed that the slope of the coastal plain shelf is 1:5000, that the rate of subsidence at the shelf margin is 2.5 cm/1000 yr, that the average sedimentation rate on the coastal plain shelf is 1 cm/1000 yr, and that sea level has remained constant for a long time. At 0.0 million years the spreading rate of a segment of mid-ocean ridge 10,000 km long changes from 2 to 6 cm/yr and stays at this rate for 70 million years, when it decreases to 2 cm/yr until 140 million years. The position of the shoreline is indicated by a dashed line. The distance from the hinge line is in 1:1 scale with the slope of the shelf.

trapezoid 500 km wide at the base, 100 km wide at the top, was elevated 5.5 km above the sea floor, and was forming from a fixed source as the plate moved past at a rate of 6 cm/yr, the displacement of water would cause sea level to rise 0.025 cm/1000 yr. Such a large feature would subside with time because of thermal contraction causing sea level to fall but not back to its original value. Formation of a single large volcanic island, of the general size of Hawaii, could displace 100,000 km^3 of water in a few million years, resulting in a change in sea level of 0.26 m. It is highly unlikely that sea level changes resulting from formation of thermal welts will ever be detected in the geologic record.

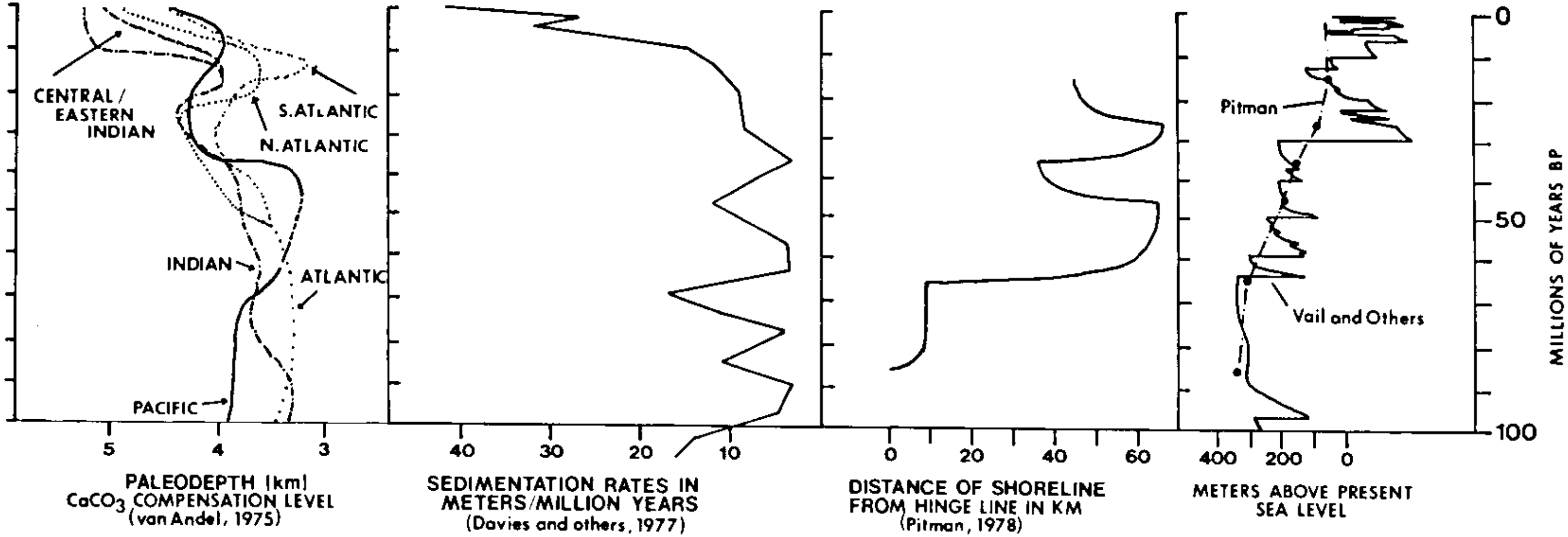

Fig. 20. Observed "sea level" curve of Vail and others (1977) compared with the sea level curve of Pitman (1978). Also depicted are the shoreline position predicted by Pitman (1978), oceanic sedimentation rate of Davies and others (1977), our recent calculation, and $CaCO_3$ compensation depth fluctuations documented by van Andel (1974). See text for further discussion.

D. *Removal of Volcanic Chains or Aseismic Ridges*

Similarly, removal of seamounts from ocean basins by subduction is a possibility, although it is more likely that subduction of linear seamount chains may be the cause of cusps in oceanic trenches and may promote the lengthening of the global trench–subduction system.

E. *Changes in Volume of Oceanic Trenches*

The global oceanic trench system along subduction zones is presently about 34,062 km long, and the trenches themselves contain about 4.99×10^6 km^3 water below 6 km of depth. Doubling the length and hence volume of the trenches below 6 km would lower sea level by only 13.9 m. From this consideration it is also evident that it is unlikely that changes in oceanic trenches have affected the sea level to an extent that could be recognized geologically.

F. *Areal Reduction of the Continental Crust*

Compression within the continents or continental collision to form mountain ranges would alter the size of the continental blocks and create more space in the ocean basins. The collision of India with Asia to form the Himalayas is a classic example (Molnar and Tapponier, 1975). Southam and Hay (1977), assuming that underthrusting of the Indian block beneath Asia and compression to form the Himalayas involved reduction of the continental block by 1×10^6 km^2, found that a drop in sea level of about 10 m would result. Pitman (1978) calculated the rate of fall of sea level during the time of collision of India and Asia to be 0.22 cm/1000 yr, or 0.15 cm/1000 yr isostatically adjusted with reference to an oceanic island.

Excluding Antarctica, continental elevations over 2000 m total 5.9×10^6 km^2 and have a volume of 10.7×10^6 km^3. If it is assumed that these elevations were produced by compression of continental material originally having an average elevation of 0.75 km (the present average continental elevation), about 3.9×10^6 km^2 of ocean basin area would be created. Assuming that the compression occurred in such a way as to add to the area of the ocean basins, with a water depth of 5.5 km, 21.5×10^6 km^3 of volume would be added to the ocean basins. This would result in lowering of sea level by 60 m. Even if all the present highlands and mountains had been produced only during the Pliocene and Pleistocene over a time span of 6 million years, the rate of sea level fall would be only 1.0 cm/1000 yr.

G. *Change in Salinity*

The effect of changes in salinity can be estimated by considering the volume of a given mass of seawater to be equal to the volume of water plus the total equivalents of salt times the apparent equivalent volume (Millero, personal communication). It is difficult to imagine how the salinity of seawater can be increased rapidly, but it is evident that during salt extraction phases, such as during the opening of the South Atlantic, salt removal may occur relatively rapidly. A salinity drop from 35 to 25‰ for the entire ocean would result in a fall in sea level of 7.6 m. In the case of the South Atlantic, a salinity change of 20‰ may have occurred in 20 million years, resulting in a lowering of sea level by 5.3 m, or 0.027 cm/1000 yr. Clearly this is too small a change to be detected in the geologic record.

H. Change in Temperature

A change in the temperature of the oceans as a whole could produce a change in sea level of 2 m for every degree Celsius (Fairbridge, 1961). Kennett and Shackleton (1976) have suggested that the thermal stratification and formation of the relatively uniform cold deep waters characteristic of the modern ocean, termed the psychrosphere, occurred 38 million years ago, during the earliest Oligocene. The average temperature of the world ocean probably dropped by 4 to 5°C in a period of only 100,000 yr, resulting in a drop in sea level of 8 to 10 m at a rate of 10 cm/1000 yr.

I. Displacement of Sediment and Change in Volume of Pore Waters

If all of the sedimentary pore space were dry, sea level would rise 489 m, which reduces to 342 m after isostatic adjustment. Obviously this is not a realistic possibility. The 1605×10^{21} g of sediment of Mesozoic–Cenozoic passive margins, active margins, slopes and rises, and pelagic sediments were mostly offloaded from the land areas of the continents and now have the effect of displacing ocean water, loading the ocean floor and continental margin and including water in pore space. If the offloading were instantaneous and no pore space were included, sea level would rise 1.65 km. After isostatic adjustment, this would be a rise of only 300 m. Including water-filled pore space in the sediment does not change the sea level as observed from land but does change the position of the ocean floor and hence the depth of the ocean. This has been modeled for 200 million years ago and is discussed in greater detail above as one of the effects of the breakup of Pangaea.

Another way of considering the effect of the sediment supply to the ocean is to recall that the total solid and dissolved load of rivers is about 250×10^{14} g/yr. Assuming that the annual sediment output is equal to the river input, 9.3 km^3 of ocean water are displaced by sediment each year. This corresponds to a rise in sea level of 2.6 cm/1000 yr before isostatic adjustment. Because this solid phase has a density of 2.7, the rise in sea level, taking isostatic adjustment into account, is only 0.5 cm/1000 yr.

J. Formation of Continental Glaciers

The glacial–eustatic sea level changes have been discussed in detail by many authors. One of the best recent discussions has been by Bloom (1967) and need not be repeated here. The characteristic of glacial–eustatic changes is that the sea level rise and fall is both large and rapid, operating on a time scale shorter than that of the isostatic response. Hay and Southam (1977) have discussed the effects of glacial–interglacial sea level changes on shelf hypsography.

K. Causes of Sea Level Changes—Summary

Table X summarizes these possible causes of sea level change, giving a rate of change and probable maximum fluctuation that could have occurred during the Mesozoic–Cenozoic. The rates of change marked with an asterisk may exceed the rate of isostatic response.

TABLE X
Possible Causes of Sea Level Change

	Sense of Change	Maximum Rate of Change Isostatically Uncompensated	Isostatically Adjusted Maximum Change During Mesozoic–Cenozoic	Duration
Addition of juvenile water	+	<1 cm/1000 yrs	154 m	Continuous
Subduction of pore water	−	<1.4 cm/1000 yrs	261 m	Continuous
Expansion of earth	−	<.18 cm/1000 yrs	275 m	Continuous
Change in volume of mid-ocean ridge system	±	<0.97 cm/1000 yrs	350 m	10^7–10^8 yrs
Change in volume of ocean basins from continental collision and compression	−	<0.22 cm/1000 years	42 m	<2 × 10^7 yrs
Change in salinity of ocean as a whole	±	<10 cm/1000 yrs	7 m	<2 × 10^7 yrs
Change in temperature of ocean as a whole	±	<10 cm/1000 yrs*	7 m	
Displacement of ocean water by sediment	+	<2.6 cm/1000 yrs*	300 m	Rate varies on scale of 10^6 yrs
Glacial eustatic rise and fall	±	<1000 cm/1000 yrs*	100 m	<10^4 yrs

12. Summary and Conclusions

Previous estimates of the masses and volumes of Phanerozoic sediments assumed the permanence of continents and ocean basins and the antiquity of the deep-sea floor. New estimates, based on new and refined data, and taking into account the theory of plate tectonics, suggest that about two thirds (1605×10^{21} g) of all Phanerozoic sediments (2485×10^{21} g) are of Mesozoic–Cenozoic age and of these about half (809×10^{21} g) reside in passive continental margins formed by the rifting of Pangaea. Mesozoic–Cenozoic geosynclines contain about 503×10^{21} g, and the pelagic sediments of the ocean basins, which are exclusively Mesozoic–Cenozoic in age, have about half the sediment of the younger geosynclines or 249×10^{21} g. The surviving Paleozoic sediments are concentrated in cratonic areas (356×10^{21} g) and in older geosynclines (524×10^{21} g).

There is general agreement among authors on the composition of the major sedimentary reservoirs, except for the pelagic sediments, where assumptions of either equal or different sedimentation rates for the biogenous components and red clay produce quite different estimates. The greatest uncertainty lies in estimating the composition of the sediments of passive continental margins which may contain larger proportions of carbonate and evaporites than is generally assumed.

Sedimentation rates in the world ocean appear to change significantly with time; during the Aptian–Albian, Campanian–Maastrichtian, Middle Eocene, and Late Miocene–Quaternary, overall sedimentation rates were about an order of magnitude higher than during the intervening periods. Organic carbon sedimentation rates vary synchronously and in the same sense, but through two orders of magnitude so that the global sedimentation system operates to greatly favor removal of carbon as organic carbon during times of high global sedimentation rate. The relatively meager evidence available from continental margins tends to confirm the global sedimentation patterns suggested by deep-sea sediments.

From exploration of the dynamics of the Earth's sediment system as a series of reservoirs and fluxes, it becomes evident that a global sedimentation system involving erosion of material from the continents, sedimentation in the deep sea, subduction, and return of material to the continents by formation of volcanic arcs cannot be in balance with regard to the masses of major chemical elements if pelagic sediments have always had a composition rich in carbonate, such as has been the case during the past 100 million years. Balance can be achieved if the older pelagic sediment had a composition similar to that of red clay. The formation of sediments and sea salts by weathering of igneous rocks cannot be regarded as a closed system, as has been generally assumed in attempts to investigate the weathering of the Earth's crust, because of the loss of pelagic sediment from the system through subduction. Examining the age and masses of cratonic and geosynclinal materials suggests that growth of continents has proceeded at an accelerating rate; the average rate of growth was 37.7×10^{14} g/yr during the Precambrian, 84.6×10^{14} g/yr during the Paleozoic, and 246.5×10^{14} g/yr during the Mesozoic–Cenozoic.

The breakup of Pangaea through rifting and separation of the continents has special implications in terms of unique sediment sequences on the passive margins. The initial rift valleys were probably occupied by stratified freshwater lakes due to the equable Mesozoic climate, and during this phase might have accumulated about 7.5×10^{21} g of organic carbon. This would be 14% of the Earth's total organic carbon concentrated on only 0.3% of the Earth's surface. The sediments rich in organic carbon are expected to be typically overlain by evaporites. It is estimated that more than half of the Earth's known evaporites were deposited in the rifts after they subsided to sea level. After separation of the continental masses, development of barrier reefs along the continental margins was typical, at least for the Atlantic. The barrier reef growth and carbonate accumulation on the continental margins became much restricted about 100 million years ago, at the same time when oceanic plankton began to remove $CaCO_3$ from the ocean waters to form calcareous oozes. Subsidence of the shelf along the continental margin generally follows the curve of subsidence of the ocean floor as it moves away from the midocean ridge, and the balance between sedimentation on the shelf and on the slope and rise appears to be a complex function of declining subsidence rate with time, varying sediment supply rate, and sea level.

Finally the possible causes and rates of sea level change have been explored in an attempt to evaluate the effects of such changes as they result from the sedimentation system and as they affect it. Possible important long-term processes affecting sea level are addition of juvenile water and subduction of pore water, which operate on the same time scale but have the opposite sign, and the expansion of the Earth. Change in the volume of the midocean ridge system has probably produced the general rise and fall of sea level during the Cretaceous and Tertiary. Displacement of ocean water by sediment has probably also had a significant effect in the Mesozoic–Cenozoic. Continental collisions and compression may be responsible for sea level fluctuation on the scale of a few million years. Changes in temperature and salinity of the ocean as a whole have probably not affected the sea level to an extent that could be detected geologically. The most rapid significant changes of sea level are those generated by the formation and melting of ice caps.

The implications of the theory of plate tectonics are pervasive and must be taken into account in any consideration of global weathering and sedimentation.

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